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MICROCLIMATE

AND

LOCAL CLIMATE

Admitted by the Ministry of Higher Education of the
USSR as a Textbook for Use in Hydrometeorological
Institutes and Universities.

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ANNOTATION

In this book an attempt is made for the first time to bring together in a systematized manner all the modern data pertaining to microclimate and local climate. The physical foundations of microclimate and local climate are analyzed, and the basic characteristics of the microclimate of the field, the forest, the valleys, the slopes, the climate of cities, and so on, are discussed, it being the case that particular attention is devoted to problems of agriculture. Considerable space is devoted to the description of the research methods employed, and to the processing of the observation data pertaining to microclimate and local climate.

The book is used as an official textbook by hydrometeorological institutes and faculties of geography in universities, and may also serve as a valuable manual for agronomists, foresters, and other specialists, who in their daily work, are constantly faced with the problems of microclimate and local climate.

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FOREWORD

Planned socialist economy utilizes to a maximum degree all the natural resources of the country and stimulates the development of all branches of Soviet science. The wide utilization of the climatic resources of the country, and the carrying out of undertakings for the transformation of its very long nature, on a scale unheard of in the entire history of mankind, was instrumental in the segregation into an independent discipline of the branch of climatology, which is concerned with microclimate and local climate.

The book "Microclimate and local climate" is designated as a textbook for schools of geography at universities and hydro-meteorological institutes. In addition, it can be used by agronomists, forestry engineers, and other specialists, who, in their daily tasks are constantly faced with problems pertaining to microclimate and local climate.

The agricultural angle was selected as the most important, from a practical point of view, and as the closest to the author, who has behind her 20 years of experience in the above indicated field.

Due to limitations in the size of the book, certain problems relating to local climate, which are treated adequately in general courses of meteorology and climatology (for example, breezes), are not analyzed in detail.

The book bases itself almost exclusively on domestic, pre-

dominantly Soviet research and investigations, which, in practical orientation and theoretical analysis, excel similar undertakings abroad.

Being the first Soviet textbook on microclimate and local climate, the theoretical bases of which are only now being developed, the book contains a series of controversial questions and aspects, and unavoidably suffers from a series of inadequacies, which in the future may be eliminated on the basis of the experience in their treatment by the VUZ's and of the practical utilization of research.

INTRODUCTION

Basic Concepts. Brief Historical Outline.

With the development of climatology, the concept of climate becomes differentiated, and at the present time, alongside the macroclimate, the local climate and the microclimate are segregated.

The differentiation of science, the segregation of its individual branches, is tied in with the expansion of our knowledge and promotes a deeper study of the segregated phenomena. In order that such differentiation is effective, it is necessary that the group of contemplated phenomena possess particular qualities or characteristics, which require special methods of investigation. Within this definition of particular characteristics may also fall particular scales of phenomena or objects. This, indeed, was the principle for the segregation of microbiology, which is undergoing successful development within the last decades.

The reason for the segregation of microclimate and local climate into objects of independent study, in addition to their specific characteristics stipulated by essentially different scales, in contradistinction to macroclimate, consists in that, first, specifically in the zone of microclimate and local climate runs a considerable part of the cycle of human activity, and, secondly, they are the most accessible for transformation in the proper direction.

The difference in scales for phenomena is determined by the scales of climate forming factors, as well as by the proximity of the contemplated atmospheric layers to the active surface, the source of heat and moisture, over which surface the basic method of transfer of all the characteristics of the air (heat, moisture, atmospheric pollution) - turbulent exchange - is on a scale hundreds and thousands of times smaller than in the free atmosphere.

Specifically, the scales, on which the phenomena occur, were the basis for the segregation of microclimate and local climate into a special study.

Under macroclimate, we understand climatic phenomena determinable by large-scale factors, such as: general circulation processes, the geographical latitude of a locality, its distance from oceans and seas, and macrorrelief.

The macroclimatic phenomena can be observed in their unadulterated form only outside of the sphere of perturbation, which is induced by local peculiarities of the active layer, i.e., only at a height of several tens or even hundreds of meters.

However, we can also analyze the macroclimate by observations at a height of 2 meters, deducting, so to speak, the effect of local conditions, by locating all observation stations in homogeneous conditions - in exposed level locations.

For example, the horizontal macroclimatic variations (latitudinal and longitudinal gradients) are measured in tenths of a degree per 100 kilometers, while the vertical gradients are measured in tenths of a degree per 100 meters.

Local climate is determined by climate generating factors of a smaller scale: mesorelief, vegetation masses, etc. We differentiate the local climate of the forest, the glade, the valley, the city, and the like. The characteristics of the local climate are manifested in a layer of air measured in tens and even hundreds of meters, but are diminished by height. Observations made at a height measured in meters and tens of meters, under various conditions of local climate, show vertical and horizontal gradients of temperature, humidity, wind velocity, which are many times greater than the respective macroclimate gradients. Thus, depending on local peculiarities of climate, the minimum temperature may vary by whole degrees for a distance of only several tens of meters, while the vertical temperature gradients in the same layer show a sharply pronounced diurnal cycle. The daylight superadiatic gradients are replaced by a nocturnal inversion, it being the case that on conversion into terms of 100 meter-distances the tenths of a degree become whole, or even tens of degrees.

Microclimate encompasses phenomena, which take place in an air layer up to 1.5 to 2.0 meters over the ground, and which are directly linked with the strictly local characteristics of the active surface: microrelief, character of vegetation, and the like. A distinguishing feature of the lowest atmospheric layer are the exceptionally high vertical gradients of temperature, wind, and humidity. If the air temperature gradients are converted into terms of 100 meter-distances they will be expressed in hundreds and even thousands of degrees. These phenomena are usually confined within the limits of the above indicated air layer, but they have tremendous practical significance, since in this particular layer ^{there} runs its course the greatest part of the human activity cycle, ^{there} takes place ~~in~~ the springing up and growth of vegetation; ~~and~~ it is specifically this layer that is most accessible to the influence of man. By changing the characteristics of the underlying surface, let us say, by destroying the grass cover, or by compressing the soil, we change the microclimate.

The above cited characteristics of the concepts of microclimate and local climate have no pretense at being final definitions. As yet, it is difficult to furnish precise formulations, considering the fact that even the general definition of climate is not yet finally worked out.

We are dealing with microclimate at the flower bed, at the asphalt highway, in the vast fields of the kolkhoz extending over hundreds of hectares, over the surface of the seas and the oceans. Due to its proximity to the active surface, the microclimate is subject to great horizontal variability in those cases, when the active surface is not homogeneous. A hillock, a ravine, a narrow

boundary in a field, each have their own microclimate. In a number of cases we are dealing with complexes of microclimates, as, for instance in a potato field, prior to the joining of the leaves, the microclimate of the open sections is different from the one of the sections overshadowed by the potato leaves. These two, in their complexity, form the microclimate of the potato field.

The scales of microclimatic phenomena as compared to the scales of general meteorological phenomena, which are depicted by a network of meteorological stations, separated by tens of kilometers from each other, are different to such an extent, that it became necessary to develop special apparatus and methods for their study. Local climate, by its scale, occupies an intermediate position, and in the USSR, up to a very recent date, it was combined with microclimate. However, the greater stability of local climate, as compared to microclimate, on account of which it must be reckoned with in the complex follow-up of the physico-geographic characteristics of a territory - its detailed zoning (as the planning of a city, health resort, sovkhos layout, roads, airports, etc.), was the reason for its segregation into an independent object of study. Particularly so, since in the study of the local climate, it is possible to use the general meteorological equipment, including aerological apparatus, with only a change in the method of its utilization. For the study of local climate, the stations are to be located not uniformly, as is being done for the study of macroclimate and weather, but with relation to the physico-geographic conditions, the effect of which upon the local climate is under study, it being the case that, in addition to studying

the features of the local climate, the factors on which they are dependent, such as relief, vegetation, and others, are also taken into account.

Figure 1 shows diagrammatically the change in one of the components of climate, the mean maximum temperature for July, along the meridian from 56 to 60 degrees north latitude. In macro-scale, the above change (in accordance with data from manuals and atlases) amounts to approximately 3 degrees (heavy continuous line), with the change taking place uniformly. Such a result will be registered by direct observations only in the case, when all meteorological stations along the above profile will be located in open level areas having a so-called natural cover (this, as far as possible, is the actual method of location of meteorological stations).

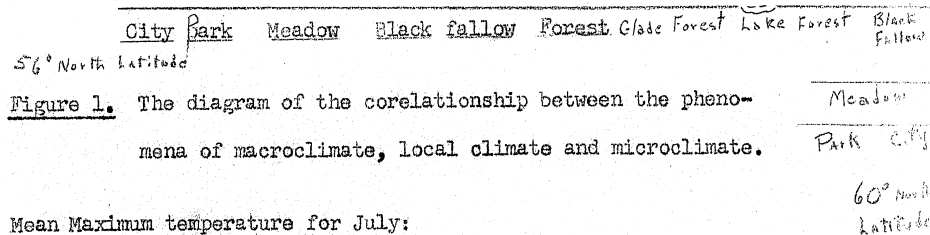
If, however, the location of the observation points will be heterogeneous, such as in a park, in a meadow, over a lake, in a city, and the observation data obtained by these stations will be used to derive the temperature change along the above profile, the matter will become considerably complicated, as shown diagrammatically by the dotted line in Figure 1.

The maximum temperature will be higher in a city, and lower in a forest, or over a lake, than in an open level locality, etc.

The variation of temperature along this profile is still more complex at a height of 5 centimeters, as represented by the

thin continuous line in Figure 1. This line will register not only the effect of the park, the city, the water reservoir, but also the effect of individual wrinkles in the soil, pebbles, and of the various denseness of the grass.

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observe them in conformity with the scales on which they occur, to locate their determinative factors.

It should be emphasized that the particular industrial possibilities opened up by the study of microclimate and local climate are determined by the complete feasibility of changing the latter in a direction that will serve our needs. This feasibility was brilliantly confirmed by the decree of the Party and the Government, which was promulgated on the initiative of Comrade Stalin on 20 October 1948 under the name "Relating to the plan for crop-protecting reforestation, for the inculcation of the grass and crop-rotation system, for the construction of ponds and reservoirs to insure high and stable harvests in the steppe - the forest - steppe areas of the European part of the USSR". The aggregate of projects, visualized by this plan, has to a considerable extent, as its specific goal, the changing of local climate and microclimate.

The peculiarities of the climate in the immediate proximity of the terrestrial surface, and the effect exerted upon them by terrestrial relief, vegetation, and bodies of water, have long since attracted the attention of scientists and practical men. However, systematic investigations of the lowest atmospheric layer began only in the second half of the 19th Century.

In 1872, on the signal tower of the Central Physical Observatory at Pulkovo, the first systematic temperature observations of the lowest atmospheric layer, at its various heights, were conducted. The results of these observations made it pos-

sible for the remarkable Russian climatologist A. I. Voyeykov to arrive at the basic conclusions on the peculiar character of the temperature stratification near the terrestrial surface. He published his conclusions in 1884, under the title "Climates of the Earth and, Particularly, ^{of} Russia". In this monograph A. I. Voyeykov deals with the law of the effect of terrestrial configuration upon the diurnal temperature range, and devotes a special chapter to the effect of vegetation upon the climate. In his conclusions with relation to the effect of local conditions, A. I. Voyeykov predominantly emphasizes the physical substantiation of phenomena and accepts, as the basis of climatic variations, the thermal balance of the active surface (radiation cycle, latent heat of evaporation, and other phenomena).

To quote one of his thoughts: "One of the most important problems of physical sciences at this time is the keeping of a book of accounts pertaining to the solar heat being received by the earth, and its air and water envelope".

This problem, in its full implications, is still with us.

A. I. Voyeykov was emphatic in his recommendations to organize special observation points and to collect field data pertaining to local characteristics of the climate.

Voyeykov's ideas were reflected and developed in the works of an expedition headed by V. V. Dokuchayev (see printed page 174).

In the volume dealing with the labors of the expedition,

devoted to meteorological observations in 1892-1894, V. V. Dokuchayev writes by way of an introduction; "The greatest number of our largest stations, by virtue of their specific location, are engaged in studying the climate of Petersburg proper, but not the climate of its surrounding marshes and wastelands; the climate of Khar'kov and Saratov but not the climate of their contiguous open steppes, the climate of Nizhniy Novgorod, Kostroma, but not the climate of the Vetluga and Unzhan forest taiga...." Thus underscoring the diversity in the climates of the city, forest, steppe, plateau, and valley, V. V. Dokuchayev points to the necessity for the studying of all these as dictated by practical considerations.

The labors of V. V. Dokuchayev and his followers have served as the basis for the greatest projects, ever undertaken anywhere in the world, in the remodeling of nature, as visualized by the Stalin plan for the overcoming of droughts and dry winds. In addition to accomplishing its basic purpose, the testing and following of the "various methods and procedures in the forest and water economies in the steppes of Russia", the Dokuchayev expedition described the effect of various planted forests upon the local climate and microclimate.

As early as 1894, G. N. Vysotskiy, who took part in the Dokuchayev expedition, characterized the basic regularities in the distribution of minimum temperatures with relation to the degree of shielding of a locality provided by the vegetation, and also with relation to the peculiarities of the underlying surface.

He followed it up in 1909 by raising the question "about local networks for the study of the special local climatic variations dependent on terrestrial relief and the character of the vegetation cover."

During the same years, G. A. Lynboslavskiy of the Forestry institute (Leningrad) studied the thermal cycle of the soil and the effect of the vegetation cover upon it.

Numerous investigations in forest meteorology were conducted by Ototskiy, Adamov, Tol'skiy, and others in experimental forest reservations.

In 1915, was published the valuable work of P. I. Koloskov "Relief as a climatic factor", assembled from the data of the Amur Regional Meteorological Bureau of the Immigration Administration.

During the First World War and the years following it, all the endeavors in this division of the climatology were directed basically along the line of accumulating factual data. Prominent in the work of this period were the investigations by V. N. Obolenskiy and A. F. Rudovits in Petrograd, G. T. Selyaninov at the Sachi Experimental Station, and A. A. Skvortsov in Central Asia, with the latter concentrating on the discovery of the physical essence of microclimatic phenomena.

Beginning with the thirties, the tempestuous development of socialist construction in our land stimulated the growth of Soviet research in microclimate and local climate with a view

to its application to the needs of agricultural production, health resort construction, transportation, national defense, and other purposes. Of particular merit are the investigations linked with the detailed agroclimatic zoning of the subtropical belt of the USSR, which were conducted by the Agrohydro-meteorological Institute under the direction of G. T. Selyaninov, the work by L. A. Golubeva and others relating to the phytoclimate of agricultural field crops, linked with irrigation projects, the penetration of transpolar areas, and other practical problems. The chair of climatology of the Geographic Faculty of the Leningrad State University, under A. V. Voznesenskiy and A. A. Kaminskiy, and later on under their followers, also conducted important works on microclimate and local climate. Extensive investigations as to the effectiveness of forest strips were conducted by Ya. D. Panfilov, V. A. Bodrov, Yu. A. Byallovich, and others. In the field of health resort construction, of greatest interest is the work by N. A. Korostelev. Research pertaining to the local climate of the subtropical by its substantiality and productive effectiveness, excels similar works performed by climatologists abroad. The works of V. A. Mikhel'son and S. I. Savinov played an important part in the study of the physical bases of microclimate, particularly the thermal balance.

After the Great Fatherland War, the USSR registers a new high in the study of microclimate and local climate and their applicability to the needs of the various branches of the national economy, with more and more attention given to the physical part of the phenomena. Thus, the ideas enunciated more than 50 years ago by A. I. Voyeykov, are now attaining their ultimate development due

to the successful work of Soviet theoreticians - meteorologists, improved technical equipment, and principally, to the gigantic growth of practical demands, linked not only with the planned mastering of, but with the actual reshaping of nature, including the climate of the lowest atmospheric layer. It is particularly necessary to mention the works of the Central Geophysical observatory imeni A. I. Voyeykov (D. L. Laykhtman, M. I. Budyko, and others).

The history of the development of microclimatology and local climatology serves as a graphic illustration of the part played by practical perseverance and production, and particularly by Soviet planned preparation in the development of science.

SECTION I

THE PHYSICAL BASES OF MICROCLIMATE AND LOCAL CLIMATE

Chapter 1

The Thermal Balance of the Active Surface.

Footnote: In this and the following chapters, devoted to the physical fundamentals of the microclimate and local climate, the thermal balance of the active surface is analyzed only in relation to its effect upon the formation of microclimate and local climate. In connection therewith, some aspects of said balance are described in greater detail, others, having no direct bearing upon microclimate and local climate, are completely left out. The approximated formulas of the thermal currents that will be given in the following text, are basically meant not for computations, but for the characteristic of the ratio between the

vertical temperature and humidity gradients in the air and in the soil, on the one hand, and the current of heat and moisture turbulent exchange, etc., on the other hand.

The surface of the soil and vegetation, and also the surface of any other body, which absorbs and emits heat by radiation, controlling thereby the thermal cycle of the adjacent layers of air and soil, is called an active surface.

The difference between the natural radiation of the soil and vegetation, on the one hand, and the back radiation of the atmosphere, on the other hand, i.e., the effective radiation, practically always has a negative sign. This means that, as a result of effective radiation, the active surface loses heat. During the daylight hours, solar radiation is usually in excess of effective radiation. The resulting heat surplus on the surface of the soil and vegetation is used up for the direct warming of the vegetation, soil, contiguous air layers, and also for the sustaining of biological processes (photosynthesis) and vaporization - transpiration.

In the absence of solar radiation, direct as well as diffused, or when it is of low intensity as, for instance, during the winter days, particularly over the snow cover, which reflects the greater part of the heat from the sun, the outgoing heat of radiation is in excess of the incoming radiation heat, as a result of which the active surface loses heat, which is compensated by the emission of heat from the air and from the soil.

The active surface plays an important part not only with

relation to the thermal economy, but also with relation to the water economy. Moisture is evaporated from the active surface and precipitated on the active surface, both, by direct condensation, in the form of dew and air-frost, and by atmospheric precipitation, in the form of rain and snow.

The thermal and water economy of the active surface constitutes the basic factor in the determination of the climatic characteristics of the lowest atmospheric layer and the upper layers of the soil, since the vertical temperature and humidity gradients in these layers are in direct ratio to the inflow of heat and water vapors from the active surface into the air, or vice-versa, from the air into the active surface. In the rare cases, when the heat and water economy of the active surface equals zero, i.e., when there is no inflow and no outflow of heat and moisture, the temperature and humidity are homogeneous through the entire layer, from the soil up to a height of 2 meters and above (the adiabatic gradient in the lowest atmospheric layer is disregarded). The greater the inflow of heat and moisture, with all other conditions being equal, the greater the vertical and horizontal differentials, since the inflow of heat and moisture into the atmosphere depends on the active surface, and it varies substantially with the variations in the latter.

The in-and-out flow account of radiation heat - the radiation balance (R), which is the heat flowing from the active surface directly into the air, or vice-versa, the heat flowing from the air (a), the heat exchange in the Soil (T), and the latent heat of vaporization (transpiration) or the heat liberated in con-

densation (E), make up in the aggregate the thermal balance of the active surface.

Under the thermal balance of the active surface we understand the balance of the heat currents that pass through this surface.

Basing ourselves on the law of preservation of energy, we come to the conclusion that the thermal balance of the active surface at every given moment must equal zero, or

$$R + T + Q + E = 0 \quad (1)$$

Strictly speaking, the components of the thermal balance should also include the in-and-out flow of heat linked with the biological processes, such as photosynthesis, respiration, etc. However, the amount of heat absorbed or liberated as a result of biological processes, as a rule, rarely attains several percent of the radiation balance, wherefore we disregard it in our coarse calculations. The considerable expenditure of heat for transpiration is included in the general losses of heat through vaporization.

The distribution of the radiation heat among the remaining components of the thermal balance is determined by the correlationship of the heat conduction of the active layer (soil, water, vegetation) to the heat conduction of the air plus conditions determining the vaporization of moisture (the presence of water in a liquid or solid state), or, in a manner of speaking, the vapor conduction of the air.

The transfer of heat, water vapor, dust, and other admix-

tures in the air, as well as of the quantity of motion of a wind current, is effected, as is known, by way of ~~the quantity of motion of a wind current, is effected, as is known, by way~~ of turbulent intermixing, which plays a leading part in the formation of microclimate and local climate, and is distinguished by great variability in time as well as in space, particularly, in a vertical direction. The sharp increase in the temperature and humidity gradients of the air, the closer the approach to the active surface, is mainly determined by the diminution in the same direction of the turbulent exchange, which at the active surface itself is infinitesimally small.

The thermal balance of the active surface unifies into a single complex not only the processes of the heating and the cooling of the air and the ground but also the two important items of the moisture cycle: vaporization and condensation. The physical basis of any characteristics of microclimate and local climate, in most cases, reduces itself to the discovery of the peculiarities of the thermal and water balance of the active surface, to the exposure of the part played by the individual components of the above balance.

This is why we begin the study of local climate and microclimate with the thermal balance of the active surface, in the analysis of which we will take into account not only the effect of local peculiarities, but also the part played by general meteorological conditions. Thus, we can discover the causes for the similarity or the diversity, as the case may be, of microclimates and local climates in various climatic zones, and also account for their variability in relation to general weather conditions.

Radiation Balance.

The radiation balance is a synthesis of short-wave and long-wave radiation. The short wave radiation, or insolation, is incident upon the active surface as direct and diffused radiation, and is partly reflected from the active surface.

If the direct radiation is designated as S , the diffused radiation -- as s , and the reflective power of the active surface in relation to short-wave radiation (albedo) -- as α , then the short-wave part of the radiation balance can be presented as:

$$(S + s) (1 - \alpha)$$

The long-wave part of the radiation balance comprises radiation of the active surface r_1 and the atmospheric back radiation. The difference between these two is usually called the effective radiation r .

The total radiation balance is expressed by the following formula:

$$R = (S + s) (1 - \alpha) + r \quad (2)$$

During the night, when there is no direct and no diffused insolation, the radiation balance is composed of long-wave radiation only:

$$R = r, \quad (3)$$

it being the case that r , as was already indicated, has predominantly a negative value.

Solar radiation. Solar radiation is the leading component of the radiation balance during daylight hours. The variation in the daylight intensity of the radiation balance, in space as well as in time, is determined to a considerable degree by the variation in direct solar radiation. Not much data is available with relation to the daylight radiation balance, since it is difficult to determine effective radiation during daylight hours. We will, therefore, take advantage of more extensive data on direct insolation and its absorption, for the comparative characterization of the daylight radiation behavior within the diurnal and annual variation, in relation to astronomical factors, general meteorological conditions, and to the radiation characteristics of the active surface.

It is common knowledge that the phenomena of microclimate and local climate are linked directly with clear weather (sunshine). Microclimatic distinctions are leveled off during overcast days. This circumstance justifies the assumption that, specifically, direct radiation is the leading factor in the shaping of microclimate and local climate.

Let us, first of all, point out that the intensity of the diffused radiation is indeed, in a predominant number of cases, considerably lower than the intensity of direct radiation.

Table 1 is a comparative compilation of the mean intensities of direct solar radiation during the hours of sunshine, and those of diffused radiation without relation to cloudiness, as observed over a perennial period during the month of July in the

environs of Leningrad. The intensity of diffused radiation is only $1/4 - 1/3$ of the intensity of direct radiation. The relative increase in diffused radiation during the hours close to the rising and the setting of the sun has no practical value, since, during those hours, both direct and diffused solar radiation are weak, and are overlapped by effective radiation.

Note: In this case, the part played by diffused radiation is analyzed only with relation to the shaping of micro-climate and local climate. When computing the ultimate value of the thermal economy, particularly in zones of predominant overcast, the part played by diffused radiation is very substantial, and it must not be disregarded.

[See next page for Table 1]

Diffused radiation is not only insignificant in intensity, but also varies little in relation to local conditions, as, for instance, it practically does not depend on the angle of slope of the active surface. Since the source of diffused radiation is the entire firmament, the variegation caused by recurring shade in the case of direct radiation does not pertain to diffused radiation.

As a result, in the case when only diffused radiation is in being, the thermal currents flowing into the ground and into the air are smaller and considerably more homogeneous in space.

TABLE 1

Intensity of direct and diffused solar radiation upon a horizontal surface in Pavlovsk in July (in cal/cm²min)

Radiation	HOURS																	
	0300	0400	0500	0600	0700	0800	0900	1000	1100	1200	1300	1400	1500	1600	1700	1800	1900	2000
Direct (s)	0.01	0.06	0.16	0.30	0.46	0.64	0.78	0.91	0.97	1.00	0.97	0.90	0.77	0.62	0.44	0.28	0.14	0.06
Diffused (s)	0.01	0.03	0.07	0.10	0.13	0.16	0.20	0.23	0.24	0.25	0.24	0.22	0.20	0.16	0.13	0.10	0.07	0.03
s/s	1	0.50	0.44	0.33	0.28	0.25	0.26	0.25	0.25	0.25	0.25	0.25	0.26	0.26	0.30	0.36	0.50	0.50

Direct radiation induces not only a sharp increase in the heat inflow, but also heat differentiation, which is due to the different angles of incidence of the sun's rays, and also to recurring shade, thereby promoting the differentiation of microclimatic peculiarities vertically, as well as in a horizontal plane.

Therefore, the evaluation of the inflow of, specifically, the direct solar radiation is absolutely necessary in the study of and the physical substantiation of the characteristics of microclimate.

The meridional intensities of insolation upon a surface perpendicular to the sun's rays vary comparatively little in space and in time (in the course of the year), since the increase in the height of the sun, and, together with it, the decrease in the path traveled by the ray through the atmosphere, observed during the transition from winter to summer [winter equinox] and from high latitudes to low ones, is compensated by the accompanying diminution in the transparency of the atmosphere due to the rise in absolute humidity.

[See next page for Table 2]

Table 2 illustrates that on a clear summer day the amount of heat obtained by an active surface, which is perpendicular to the sun's rays, is practically of the same order of magnitude from Parallel 40 to Parallel 80 north. In other words, over the

TABLE 2

The annual variation in the meridional intensities of direct insolation upon a perpendicular surface (in cal/cm²min)

Observation Point	ϕ	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
Tikhaya Bay	80°	-	-	0.68	1.11	1.20	1.25	1.27	1.16	0.90	-	-	-
Polyarnoye	69	-	0.55	0.97	1.25	1.30	1.22	1.03	1.10	1.16	0.90	0.68	-
Leningrad (Pavlovsk)	60	0.92	1.06	1.22	1.26	1.27	1.24	1.21	1.22	1.20	1.14	0.98	0.79
Moscow	56	0.92	1.06	1.18	1.23	1.22	1.19	1.22	1.18	1.17	1.10	0.96	0.80
Alma-Ata	43	1.25	1.29	1.32	1.27	1.29	1.27	1.27	1.24	1.22	1.28	1.25	1.19
Tashkent	41	1.26	1.33	1.33	1.31	1.29	1.28	1.28	1.27	1.26	1.22	1.24	1.25

entire territory of the Soviet Union, from Cape Molotov on the Island of Severnaya Zemlya to the southern boundaries of Tadzhikistan, may be assorted any number of sloping surfaces, onto which the meridional inflow of direct insolation on a clear summer day will be the same. The annual variation of the intensity of insolation upon a surface perpendicular to the sun's rays is also small, with the exception of protracted polar night zones.

The intensity of insolation upon a horizontal surface varies considerably even during the summer. In the winter, the variations with latitude are very sharp. (See Table 3).

Table 3

The annual variation of the meridional intensities of direct insolation upon a horizontal surface (cal/cm²min)

Observation												
Point	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
Tikhaya Bay	-	-	0.09	0.40	0.56	0.68	0.65	0.48	0.20	-	-	-
Polyarnoye	-	0.11	0.40	0.60	0.84	0.85	0.78	0.62	0.47	0.19	0.07	-
(Leningrad												
(Parloysk)	0.07	0.24	0.43	0.69	0.91	1.00	1.05	0.96	0.80	0.57	0.30	0.14
Alma-Ata	0.54	0.72	0.93	1.06	1.17	1.19	1.18	1.08	0.93	0.79	0.59	0.47

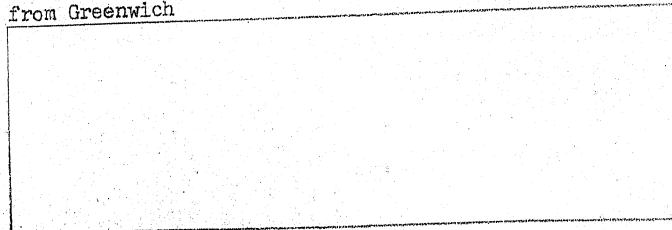
During the summer, the intensity of direct insolation upon a horizontal surface in Leningrad is only by 20 percent lower than in Alma-Ata. Therefore, with all other conditions being equal, the

thermal inflow into the air, and consequently, the microclimatic features, specifically, the vertical temperature gradients on a clear midday should be not much lower in Leningrad than in Central Asia. But in Leningrad the number of clear days is considerably below that in Central Asia. The synoptic map in Figure 2 shows that in the deserts of Central Asia the non-overcast days (clear and mild sky cover) make up 95 percent, in Leningrad -- 50 percent, and in the north of the Kola Peninsula (Station Polyarnoye) -- only 25 percent of all days.

Hence the conclusion that, although on individual days in the north, the intensity of insolation may induce the formation of a clearly pronounced microclimate, such days are the exception, while in the south they are predominant.

Figure 2 Recurrence of non-overcast skies (cloud formation indicated by 8 balls) in July

from Greenwich



The technical difficulties of direct observations of the elements that make up the radiation balance, including direct insolation, force us to resort to indirect data by utilizing the link between radiation and the more accessible to observation

meteorological and astronomical elements.

In approximate computations, direct insolation can be considered a function of only two variables: the solar altitude and the degree of sky cover. The solar altitude can always be determined by the latitude, season of the year, and time of day (See Appendix 3)

The link between the intensity of direct insolation upon a horizontal surface S , when the solar altitude h_0 (with $h_0 > 10^\circ$) for the temperate latitudes is close to rectilinear, as shown in Figure 3, which furnishes data observed at Pavlovsk, Uellen, and Potsdam.

Small Calory
CM²/MIN

Figure 3 The functional relationship between direct insolation S upon a horizontal surface and the solar altitude h_0 , as per observations made at Uellen (1), Potsdam (2), and Pavlovsk (3)

If S and h_0 are expressed in percentage of their maximum values (in practical application, these are their meridional values), the link between S and h_0 can be presented mathematically as follows:

$$S = 1.1h_0 - 10 \quad (4)$$

This formula may be used beginning with a solar altitude of 10 degrees, for the approximate characterization of the diurnal variation in insolation intensity.

The degree of sky cover can be presented visually by the sign in the scale number evaluation of the latter, or fixed by the heliograph. The generally accepted grouping of observations in the lowest atmospheric layer by the month, season of the year, hours of the day, and by degree of sky cover (clear, cloudy, and overcast), is, in essence, the distribution of data relating to the intensity of insolation.

Since it is not the cloudiness itself, but rather the degree of the covering of the sun caused by it, that affects direct insolation, it would be better to substitute the division by cloudiness by a division into the following groups: with sunshine, with varying sunshine, and without sunshine.

Keeping track of direct insolation is of great interest for the additional reason that it can vary substantially as a result of the geometric peculiarities of the underlying surface, which, in the final analysis, leads to a great variegation in the radiation balance. The different sides of a mound, of a vegetation stem, of a tree trunk, of a tree, of the ridge and of the furrow of a plowed-up field, of the slopes of hill and valley, all these receive a different amount of direct insolation, in relation to the orientation and the degree of steepness of the side or the slope. These variations are linked directly to the solar altitude and azimuth (i.e., with the latitude, season of the year, and

time of day), and they can be computed with aid of simple trigonometric formulas.

A. F. Zakharova computed the flow of direction insolation onto southern and northern slopes, 10, 20, 30 and 40 degrees steep for latitudes 42, 50, 60 and 70 degrees north. In analyzing the data, A. F. Zakharova came to the following conclusions:

During the winter season, between the days of the autumn and spring equinoxes, the diurnal variation in the intensity of direct insolation upon slopes of various exposure, is regular and monotypical for all latitudes within the boundaries of the USSR. On the southern slopes, the intensity of insolation at all hours increases with the steepness of the slope, on the northern slopes it increases, with the maximum differentials observed during the hours around midday.

During the summer period, particularly around the day of the summer solstice (22 June), the diurnal variation in insolation intensity is considerably more complex. The latter is due to greater solar altitude and the resulting increase of the solar ray's vertical component, the projection of which upon a surface normal to the slope depend only on the steepness of the slope, and will be the smaller, the greater the steepness, regardless of its exposure. In addition, up to 0600 hours and after 1800 hours, the disposition of the southern and northern slopes with relation to the sun changes as compared with their daylight disposition: the sun is in the northern part of the firmament, lighting up predominantly the northern slopes, and sets earlier

over the southern slopes than over a horizontal surface, the earlier, the steeper the slope. The first circumstance is of predominant importance in low latitudes (50 degrees north and to the south of it), the second circumstance is of predominant importance in the north, particularly in the transpolar regions.

During the summer, to the south of Parallel 50 north, the intensity of direct insolation on the southern slopes in the midday hours is almost no different from its intensity on a horizontal surface, and in the subtropics, the steep southern slopes receive even less heat. In the morning and evening hours, at all latitudes, but particularly in the north, the southern slopes receive less insolation than the northern slopes. In the Transpolar basin, with 'round the clock daylight, two maximum insolation intensities are received by the steep northern slopes: at 0400 hours and at 2000 hours; on the steep southern slopes, with a sharply pronounced insolation maximum during the midday hours, the sun actually "rises" only at 0400-0500 hours and "sets" at 1900-2000 hours.

In Table 4 (as per A. F. Zakharova) are cited the diurnal summations of heat from direct insolation received by the northern and southern slopes of various steepness during the days of the summer and winter solstices and the vernal and autumnal equinoxes at Parallels 50 and 60 north (in clear weather). For comparison purposes, insolation data in relation to a horizontal surface are included.

The maximum variations are observed on the days of the vernal

and autumnal solstices. On these days, a northern slope 40 degrees steep is in the shade during the entire day, while the southern slopes, the steeper they are, the more insolation they receive. A southern slope 40 degrees steep, in the diurnal summation, receives 200 cal/cm² more than a northern slope.

A southern slope 15 degrees steep at Parallel 60 (the latitude of Leningrad) receives on a clear day as much insolation as a horizontal surface at Parallel 50 (the latitude of Khar'kov). On the other hand, a northern slope, 10 degrees steep, at Parallel 50, on a summer day, receives even less direct insolation, than a horizontal surface at Parallel 60.

Table 4

Diurnal summations of heat received by northern and southern slopes through direct insolation (in cal/cm²)

Data	Northern Slopes				Horizontal Surface	Southern Slopes			
	Steepness					Steepness			
	40	30	20	10		10	20	30	40
$\phi = 50^\circ$									
22/VI	441	551	630	686	727	745	739	721	678
21/III 23/IX	0	101	197	289	376	443	505	547	577
22/XII	0	0	0	27	88	136	193	235	281
$\phi = 60^\circ$									
22/VI	388	488	575	642	709	716	733	727	705
21/III 23/IX	0	0	91	186	270	346	412	464	503
22/XII	0	0	0	0	14	41	67	89	112

In Table 5 are cited the diurnal summations of heat received by the southern slopes in various latitudes and expressed in percentage of the summation of heat received by a horizontal surface at latitude 42 degrees north in clear weather (as per A. F. Zakharova).

On the day of the summer solstice, the southern slopes in all latitudes receive somewhat less heat than a horizontal surface in the subtropics, and, generally, latitudinal variations are small. During the vernal and autumnal equinoxes, all the way up to the latitude of Leningrad, steep southern slopes receive more heat than a horizontal surface in the subtropics.

Table 5

Summations of heat received by southern slopes in percent of summations of heat received by a horizontal surface

Data	Latitude																	
	$\phi = 42^\circ$				$\phi = 50^\circ$				$\phi = 60^\circ$				$\phi = 70^\circ$					
	Steepness																	
	0	10	20	30	40	10	20	30	40	10	20	30	40	10	20	30	40	
22/VI	100	101	98	92	83	100	99	97	91	96	99	98	95	92	95	96	95	
21/III	23/IX	100	114	125	131	132	98	112	120	127	76	91	102	110	50	64	76	88

If it is assumed that the basic factor, determining the azonality of the southern and northern slopes, is the inflow of direct insolation,

then it must be recognized (proceeding from Table 4) that the northern slopes are azonic in all latitudes throughout the year, while the azonality of the southern slopes, in relation to the inflow of direct insolation during the summer, is leveled off, particularly in the low latitudes.

In the interlaced structure of terrestrial relief, the obstruction of the horizon to the oppositely disposed slopes has a great effect upon the inflow of direct insolation.

Figure 4 shows the obstruction of the horizon in a canyon at two closely located stations, in a mountainous zone, at latitude 39 degrees north. The internal boundary of the crosshatched part of Figure 4 corresponds to the height (in degrees) of the line of the horizon. At the northern slope station, naturally, the horizon is obstructed to a greater extent from the south side, at the southern slope station -- from the North side. But even the opposite part of the horizon is obstructed to a considerable extent (in places, up to 30 degrees). As a result, the factual rising and setting of the sun lag behind (see solar path for 22 June and 22 March), and in the winter (22 December), even at the southern slope station, the sun rises for a very short time after midday only, while at the northern slope station it does not show up at all.

Since the slope exposure basically affects the inflow of direct insolation, the final effect is linked to a considerable extent with the recurrence of clear weather, and will, therefore, manifest itself with particular sharpness under continental conditions.

Figure 4. Obstruction of the horizon at the Varzobsk Botanical Garden meteorological stations (I -- northern exposure station; II -- southern exposure station).

We will yet return to the part played by the radiation cycle in the formation of climates on slopes while now we turn to the analysis of other characteristics of the active surface, which also affect the inflow of insolation.

Vegetation retards direct insolation to a considerable extent. According to observations by Sakharov, only 5 percent of total insolation penetrates into thick whortleberry - barberry fir tree forest (See Figure 5). The intensity of insolation is also diminished in the depth of high grass. Table 6 shows results of observations by A. M. Kutyreva in the Leningrad base of the Agrohydrometeorological Institute.

The percentage of penetration by the sun's rays into the depth of the grass cover depends to a high degree on the density of the latter, disposition, shape and compactness of the foliage. These characteristics basically determine the diversity of microclimates in the midst of herbaceous vegetation.

It should be noted that the penetration of the sun's rays depends on the solar altitude. The lower the solar altitude, the smaller the percentage of rays that penetrate into the depth of the herbaceous vegetation cover. This feature complicates the diurnal variation of microclimatic phenomena.

Figure 5. Radiation in forest phytocycoses, expressed in

percentage of radiation incident upon an exposed surface (M. I. Sakharov). 1 - exposed surface; 2 - lichen-moss pine forest; 3 - sphagnum pine forest; 4 - red bilberry bush pine forest; 5 - raspberry bush pine forest; 6 - hemp nettle oak forest; 7 - hemp nettle pine forest; 8 - hazelnut pine forest; 9 - whortleberry bush pine forest; 10 - spiraea alder tree forest; 11 - green moss pine forest; 12 - whortleberry-barberry fir tree forest.

With relation to evaluating conditions of hibernation of winter crops, the degree of penetration of solar radiation into the snow cover is of great interest. According to observations by N. N. Kalitin in Pavlovsk, 20 percent of insolation absorbed by the snow cover penetrates to a depth of 10 centimeters. According to Ol'son, 50 percent of solar radiation penetrates to a depth of 10 centimeters, and 10 percent to a depth of 30 centimeters.

Not all the insolation incident upon an active surface is absorbed by the latter. Part of the radiation, depending on the reflecting capacity of the surface, is returned. The reflecting capacity of various surfaces is given in Table 7.

Table 6

Intensity of total radiation in standing wheat ("Lutescence" 062 with "Novinka" on 20 August 1937 at 0900-1000 hour

Height, in cm	cal/cm ² min	Percent
150	0.81	100
70 (2/3 growth)	0.70	86
35 (1/3 growth)	0.46	57
9 (meter on the soil)	0.20	25

Table 7

The reflection factor (albedo) of some surfaces

Surface	Albedo (in percent)
Fresh snow	85
Snow mixed with dirt and melted	40
Denuded soil, light (quartz sand)	35
" " dark, dry	15
" " moist	10
" " freshly plowed	5
Light grass, cereals in first stages of ripening	25
Dark grass, cereals in first phases of growth	20
Red bilberry bushes	10
Deciduous forest	20
Fir tree forest	10-15
Water (for direct radiation)	
with solar altitude 45°	5
" " " 15°	20
" " " 5°	55

The snow cover has the maximum albedo. The albedo of melted snow, particularly when polluted with dirt, is sharply diminished. Freshly plowed black soil has the lowest albedo.

Data on the albedo of winter and summer crops, with relation to the phase of growth, were obtained by B. N. Gal'perin,

as per observations of Southeastern Grain Crop Institute, USSR. During the first periods of growth, when the grass does not yet cover the entire soil, the albedo of the field hardly attains the value of 16 percent. With the further growth of the crops, the albedo is increased, and, to the moment of ripening, it attains the value of 24 percent.

The same author reports that the "multuring" of the summer wheat sowings with straw considerably increases the albedo of the field, particularly during the first period of growth. Non-"multured" field, during this period, has an albedo of 10 percent, while a "multured" field has an albedo of 23-24 percent.

The albedo value is one of the large factors that go into the determination of the varieties in microclimate and local climate. It has long since been known that in the environs of large cities the snow disappears one-two weeks sooner than in places far away from the latter. This phenomenon becomes easy to understand when we compare the amount of insolation absorbed by the polluted snow cover of urban area fields (albedo 40 percent), which, with an intensity of $1 \text{ cal/cm}^2\text{min}$, will equal $0.6 \text{ cal/cm}^2\text{min}$, with the amount of insolation absorbed by the pure snow (albedo 85 percent) of the fields far away from industrial plants and transportation equipment, which, with the same intensity of insolation, equal only $0.15 \text{ cal/cm}^2\text{min}$.

T. G. Berlyand constructed albedo charts for the European territory of the USSR, by the season of the year. The winter and summer albedo charts represented in Figures 6 and 7 must, of course,

be considered as first approximations only. During the winter, an albedo minimum value is observed in the south, where the snow cover lacks stability. The second albedo minimum is observed in the forest belt, where the trees are only partly covered with snow. It is possible that the mean albedo value of an under-snowed forest is lower than the one given in the chart. During the summer, the minimum albedo value is also in the forest zone. One albedo value maximum is adapted to the Caspian desert zone, and the second maximum is adapted to the tundra.

Due to great reflecting capacity, the seasonally small amounts of insolation, under the conditions of our snow-pre-dominant winters, lose all practical value, and the decisive importance at night, as well as during the day passes to the long-wave radiation -- effective radiation.

Figure 6. Albedo (in percent) during the winter season on the European territory of the USSR (according to T. G. Berlyand).

Figure 7. Albedo (in percent) during the summer season in the European territory of the USSR (according to T. G. Berlyand).

Effective radiation. The analysis of the physical characteristics of the processes, by which effective radiation is stipulated, is given in general text books on meteorology. We will limit ourselves here to a direct analysis of the effective radiation proper and its relation to meteorological conditions.

Effective radiation is tied indirectly with the temperature of the radiating surface. This circumstance, incidentally, somewhat levels the variations in the radiation balance of various surfaces. The increased income in short wave radiation, which may be due to a reduced albedo, and, all other factors being equal, the increased temperature, resulting from it, leads to an increase in effective radiation.

It was established both experimentally and theoretically that effective radiation can be determined by temperature and humidity in the air and by the degree of cloudiness. There are a number of methods for the computation of effective radiation during the night, when long-wave radiation is the only component of the radiation balance. It is also the determinant of the intensity of the nocturnal chilling and of the near-the-surface temperature inversions, which, with regular temperatures close to zero, may result in radiation frost.

Figure 8 is a nomogram by I. G. Lyntershtein and A. F. Chudnovskiy, with the aid of which it is possible to compute the nocturnal effective radiation by the temperature and humidity of the air.

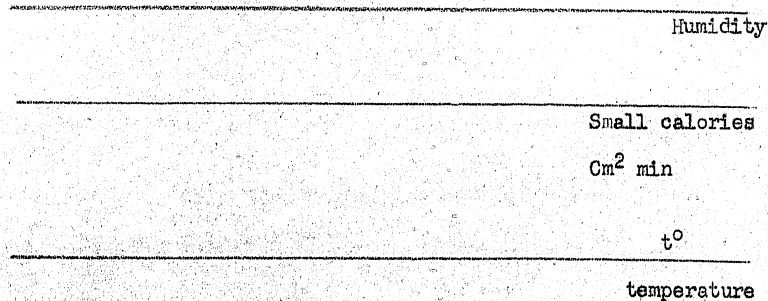


Figure 8. Nomogram for the determination of nocturnal radiation balance (effective radiation) under a clear sky (as per I. G. Lyntershtein and A. F. Chudnovskiy)

Applying the ruler in such a way that its edge is on the corresponding graduations of absolute humidity by the upper scale and temperature by the bottom scale, we lay off along the same edge of the ruler, on the middle scale, the value of the nocturnal radiation balance. It follows from Figure 8 that, assuming a constant absolute humidity, effective radiation will increase with higher temperature. Conversely, with constant temperature, effective radiation will decrease with higher absolute humidity.

However, when juxtaposing effective radiation against relative humidity, the result, upon first examination, is rather unexpected, to the effect that effective radiation, in first approximation, is almost entirely determined by relative humidity, and does not depend on temperature.

This phenomenon is explained as follows.

In order for relative humidity to remain constant irrespective of the rise in temperature, an increase in absolute humidity is prerequisite. As a result, the effect of a rise in temperature upon the increase in effective radiation is compensated by the reverse effect of the increased absolute humidity, and, in the final analysis, effective radiation remains constant.

With relation to relative humidity, effective radiation on a clear summer night oscillates around $0.1 \text{ cal/cm}^2 \text{ min}$, when the

humidity is 10 percent. A nocturnal relative humidity on the order of 10-20 percent is only possible during foehns and dry winds, and only under these conditions an effective radiation in excess of 0.2 cal/cm²min is possible.

Effective radiation depends substantially on vertical temperature variations, and, in the presence of deep temperature inversions, as, for instance, occur in Eastern Siberia, is considerably weakened.

During the summer, effective radiation has a clearly pronounced diurnal cycle, increasing toward midday. The latter phenomenon is determined not only by a sharp rise in the temperature of the active surface, but also by the sharper (than at night) drop in temperature with height. This makes all the empirical formulas for the computation of nocturnal effective radiation, as well as the Figure 8 nomogram, constructed on the basis of nocturnal observations, subject to additional corrections, when daylight effective radiation is to be computed.

Sky cover, particularly lower cloud formations, exert a decisive influence over effective radiation. Effective radiation under conditions of sky cover r_n can be expressed by the formula

$$r_n = r_0 (1 - K_1 n_1 - K_2 n_2 - K_3 n_3), \quad (5)$$

where r_0 is the effective radiation under clear sky condition;

n_1 is the amount of lower cloud formation (Cu, Sc, Ns, St); n_2 - the amount of middle cloud formation (Ac, As); n_3 - the amount

of upper cloud formation (C_1 , C_s), the degree of cloudiness being expressed in decimals; $K_1 = 0.8$, $K_2 = 0.5$, $K_3 = 0.2$ (the values of the factors K were determined by Yefimov from observations in the environs of Leningrad).

By the use of Figure 8 and formula (5) we determine that, with unbroken lower cloud formation and a 100 percent humidity, effective radiation equals $0.03-0.04 \text{ cal/cm}^2\text{min}$, i.e., 4-5 times less than under a completely clear sky.

In the presence of an overcast sky and a temperature inversion, effective radiation may even have a positive sign, i.e., as a result of a long-wave heat exchange, the active surface will not lose, but, on the contrary, will gain heat. Such phenomena are observed, particularly, in warm advections during the winter season.

With increase in the altitude of a locality, effective radiation is increased due to the diminishing of the water vapor content in the air and the attenuation of the nocturnal and winter temperature inversions. If part of the firmament is obstructed, as is the case in deep synclines and canyons, or in narrow clearings, or at the walls of buildings, effective radiation is correspondingly diminished. Tops of elevations sustain a greater loss of heat, due to effective radiation, than synclines and canyons.

The effective radiation of the ground vegetation cover under the joined canopy of a forest is negligible, since, under these conditions, the long-wave heat exchange takes place between the

ground vegetation cover and the bottom surfaces of the tree tops, with the temperatures of both close to each other. In the presence of non-joined tree tops, the effective radiation of the ground vegetation cover is, naturally increased with the increase in the gaps between the tree tops. All these features must be taken into consideration when the factors determining microclimate and local climate are studied.

The relationship between effective radiation and the degree of cloud formation and temperature determines the variability of effective radiation on a macroclimatic scale, particularly during the summer. The decrease in the number of overcast days and the parallel temperature rise evident when moving from the tundra belt to the southeastern deserts, result in an average increase of effective radiation intensity of 70 percent. But on individual clear nights, also in the north, there is, of course, possible a great loss of heat through radiation chilling.

Radiation balance. The direct effect on the thermal processes in the lowest atmospheric layer is exerted not by individual components of the radiation balance, but by the radiation balance as a whole, which is the algebraic sum of all its components. It is specifically the radiation balance, which goes into all quantitative computations, where radiation is involved.

The determination of periods with positive and negative radiation balances, in both its diurnal and annual cycle, is of great practical interest, since the sign of the radiation balance, in the majority of cases, determines the sign of the temperature and humidity gradients, i.e., it indicates the direction of the

thermal current, and also the presence of vaporization--transpiration, or, conversely, of water vapor condensation.

The Lauberer observations in Central Europe, later confirmed by special investigations under various climatic conditions of the USSR, conducted by the Central Geophysical Observatory, have shown that, during the diurnal cycle, the transition from a negative to a positive radiation balance and vice-versa, occurs not at the rising or setting of the sun, but when the altitude of the sun is approximately 10-15 degrees.

As an example, Figure 9 presents the data collected by I. G. Lyutershtein at the Tashkent Observatory.

Figure 9. The time of the zero values of the radiation balance as observed by I. G. Lyutershtein in Tashkent ($\phi = 40^\circ$), with relation to solar altitude.

1 - $R = 0$; 2 - $h_\odot = 10^\circ$; 3 - $h_\odot = 15^\circ$

In the morning the zero radiation balance occurred at a solar altitude of somewhat less than 10 degrees, and in the afternoon -- at 10 - 15 degrees. These moments, as will be seen further below, separate in time the part of the diurnal period characterized by adiabatic and superadiabatic temperature gradients, by intensive vaporization-transpiration, by considerable turbulent intermixing from the part of the diurnal period characterized by temperature inversion, by air stagnancy, by possible moisture condensation in the form of dew or fog.

Table 8 below shows the mean time for the ante-meridian (A) and post-meridian (B) passing of the radiation balance through its zero value for the 15th of each month. This table cannot be used during the presence of the snow cover, since, due to the higher albedo of the snow, the positive balance is effected at a considerably greater solar altitude (20-25 degrees).

TABLE 8

The time of establishment (A) and discontinuance (B)
of the positive radiation balance of an active
surface (expressed in hours)

Months	Geographical latitude					
	40°		50°		60°	
	A	B	A	B	A	B
IV	0500-0700	1600-1800	0500-0700	1600-1800	0500-0700	1600-1800
V	0500-0700	1700-1900	0400-0600	1700-1900	0400-0600	1700-1900
VI	0400-0600	1700-1900	0400-0600	1800-2000	0400-0600	1800-2000
VII	0500-0700	1700-1900	0400-0600	1700-1900	0400-0600	1700-1900
VIII	0500-0700	1700-1900	0500-0700	1600-1800	0500-0700	1700-1900
IX	0500-0700	1600-1800	0500-0700	1600-1800	0600-0800	1500-1700
X	0600-0800	1500-1700	0600-0800	1500-1700	0700-0900	1400-1600

The variation in the intensity of the radiation balance during the daylight hours runs parallel with the variation in solar altitude, while the nocturnal radiation balance shows little variation in time, provided the degree of cloud formation is the same.

Figure 10 shows the diurnal cycle of the radiation balance under the conditions of summer clear weather over a dry bottomland meadow in the environs of Leningrad (Koltushi) and over a demided surface in Tashkent.

cal/cm²min

Hours

Figure 10. Diurnal cycle of the radiation balance.

1 - mean value at Tashkent for June as per I. G. Lyutershtein;
 2 - same for Koltushi on a clear day in July; 3 - same for Koltushi
 on an overcast day in July.

It is notable that, regardless of the lesser solar altitude during the observation (in Koltushi the midday = 50°), the intensity of the radiation balance of the meadow, under a clear sky, is at all daylight hours, greater than the same for Tashkent (midday = 70°). This is primarily due to the considerably higher degree of effective radiation under conditions of the desert, where the temperature of the surface of the soil was in excess of 50 degrees Centigrade for a few hours and, secondarily, to the considerably higher albedo of the light-colored forest soil, amounting to 35 percent, as compared with the albedo of the grass cover (amounting to 20 percent).

Figure 10 illustrates another fact, which is of great importance in the general thermal economy. The intensity of the negative nocturnal radiation balance is 5 to 6 times lower than the intensity of the positive daylight radiant balance. Such a correlation between the daylight and nocturnal intensities of the radiation balance is characteristic for the summer season.

The diurnal variation of the radiation balance is linked basically to the variation in direct solar radiation. Special juxta-positions of the radiation balance variations with solar altitude show the possibility of using the above cited formula (4) also for radiation balance computations.

In its first approximation, the diurnal variation of the daylight radiation balance may be computed by the solar altitude, as follows: the moments of zero radiation balance are determined by the solar altitude of 10 to 15 degrees (the center of intervals cited in columns A and B, Table 8). The variation in the radiation balance during the intermediate hours is determined in percentage of the midday balance of the variations in the solar altitude, which is also expressed in percentage of the meridional solar altitude. (See formula 4).

Such computations may be useful in the extrapolation of the results of direct observations of the radiation balance at midday for the remaining daylight hours, in the presence, of course, of stable clear weather.

The data for one of the overcast days in July, which is also presented in Figure 10, visually illustrates the decisive effect of cloud formation upon the intensity of both the daylight and nocturnal radiation balance.

The effect of cloud formation on the nocturnal radiation balance was already analyzed in the section devoted to effective

radiation (see pages 27 to 28), and it reduces itself to a direct reduction in intensity with increased cloud formation and lowered cloud altitude. In daylight, the situation is more complex, since with the increase in cloud formation, direct solar radiation is attenuated, whereas diffused radiation is increased, and loss of heat through effective radiation is reduced. As a result, the maximum daylight radiation balance is observed not under conditions of a perfectly clear sky, but rather in the presence of clouds, which do not obstruct the inflow of direct solar radiation.

Figure 11 shows the variation in the mean hourly intensity of the radiation balance during the near midday hours (1100 and 1300 hours) for July and August in the environs of Leningrad, in relation to the general degree of prevailing cloud formation.

cal/cm²min

Clear Intermittent Overcast

Figure 11. Intensity of the radiation balance during near-midday hours in relation to general cloud formation.

● - July; ○ - August.

Into the interval of cloud formation, designated by 7 to 10 balls, fall both the maximum and the minimum intensities of the radiation balance with the mean intensity equal to 0.8 cal/cm²min.

With cloud formation designated by less than 7 balls, the radiation balance intensity is maintained at 0.7 to 0.8 cal/cm²min. With cloud formation designated by 0 to 2 balls, the mean radiation balance intensity is 0.75 cal.cm²min, and with cloud formation designated by 3 to 6 balls, it is 0.78 cal/cm²min.

The study of direct solar radiation, effective radiation, and the radiation balance as a whole, exposed a series of local peculiarities, which are determinable by the diversity in the radiation capacity of adjoining parcels of the active surface. But simultaneously with the above, a certain homogeneity, which, in some individual cases, particularly during clear summer days, brings into close proximity the radiation conditions of the subtropics with those of the transpolar areas, became apparent. Alongside of that, the macroclimatic variations (as in cloud formation), which determine the characteristics of the radiation cycle of the climatic zones of the USSR, and consequently, the peculiarities of the formation within these climatic zones of microclimate and local climate, also came into view. All this will be of great subsequent value in the analysis and synthesis of the vast amount of data, with relation to the climate of the lowest atmospheric layer, which data has accumulated in the course of recent decades.

Chapter 2

HEAT EXCHANGE IN THE SOIL

Radiation energy absorbed by the active surface is transformed into heat energy and is transmitted both into the depth of the active layer and into the adjacent air layers. Under the active layer in this case we understand layers of the soil with its vegetation or artificial cover (for example, asphalt), in which there are diurnal and annual thermal economies, determinable by the diurnal and annual variations in the radiation cycle. In [sizable] water basins, the active layer, naturally, is the water itself, in small ponds, the underlying bottom may be a part of the active layer.

Let us now analyze the mechanism of heat transfer in one of the most widespread active layers -- the soil.

The thermal flow into the soil is determined by two factors: the temperature gradient and the heat conduction of the soil. The greater the difference in the temperatures between the surface and the deeper layers, the greater will be the amount of heat flowing into the soil, or, conversely, flowing out of the soil.

The calculation of the thermal inflow into the depth of the soil is made by formula.

$$T = t_2 - t_1 \quad z = cK \quad \frac{t_2 - t_1}{z_2 - z_1} \quad z, \quad (6)$$

Where T is the thermal flow in cal/cm²sec; λ is the heat conduction; $\frac{t_2 - t_1}{Z_2 - Z_1}$ is the vertical temperature gradient, i.e., the mean difference in temperatures ($t_2 - t_1$) at two depths over some time interval τ , divided by the difference in the depths ($Z_2 - Z_1$). Since, under field conditions, it is hard to determine the coefficient of heat conduction λ , the latter is substituted by its equivalent, which is the product cK [volumetric thermal capacity by temperature conduction, see formula (6)].

To determine the amount of heat absorbed in the heating or liberated in the cooling of a certain layer of the soil Z , the following formula is used:

$$\Delta T = cZ (t_1 - t_0), \quad (7)$$

where ΔT is the increment of heat in calories over a definite time interval, c is the volumetric heat capacity of the soil, ($t_1 - t_0$) is the mean difference in temperatures for the entire layer Z over the same interval.

Both methods are used simultaneously. By the second method, they establish the amount of heat used up for the warming of the upper layers of the soil, the temperature of which is measured directly. The amount of heat used up for the warming of the deeper layers of the soil, for which there is no temperature data, is determined at the upper boundary of this layer with the aid of the first formula.

The above formulas are approximate, yet adequately precise for the solution of a series of practical problems. It is to be taken into account that precision in the computation of the diurnal heat exchange in the soil, with modern methods of observation (the Savinorsk thermometers), is usually limited not by the approximateness of the formula, but rather by errors in reading the temperature and in the interpolation of the latter, particularly in the layer 0 to 5 centimeters.

The coefficient of heat conduction^{sc} is numerically equal to the amount of heat flowing in one second through one square centimeter of a layer of homogeneous substance, one centimeter thick, if the temperature ~~is~~ between the top and bottom of the layer, differs by one degree Centigrade, that is, with a temperature gradient of one degree per one centimeter.

The value of the coefficient of heat conduction is determined by the physical characteristics of the soil, by which not only the so-called molecular heat conduction but a series of other phenomena are stipulated. In the transfer of heat from the surface to the depth of the soil, and vice versa, a definite part is played by the movement of water in liquid as well as in gaseous state. Water, percolating into the depth of the soil, or conversely rising by way of capillaries, carries the heat with it. The heat used up for water vaporization in one layer, may be liberated in another layer, where the water vapors are being condensed.

In Table 9 are cited the coefficients of molecular heat conduction for various soil components.

TABLE 9

Coefficients of molecular heat conduction

Soil components	cal/cm sec deg
Feldspar	0.0058
Limestone	0.0040
Peat	0.002
Ground water	0.0012
Ground air	0.00005

The heat conduction of solid particles of soil is 100 times greater than the molecular heat conduction of the air it contains. Since the physical properties of the soil, its compactness, and particularly its water and air content are subject to variations, its heat conduction will, naturally, vary.

The porosity of the soil, its structure, the nature cohesion of its particles, the state of its moisture content, greatly affect its heat conduction. Porosity is characterized by the ratio of the volume of the pores to the volume of the soil, as a whole, expressed in percent.

With an increase in porosity, the heat conduction of the soil drops, as seen from the curve in Figure 12, which was constructed from data of dry soils of various porosities. A particularly intensive drop in heat conduction occurs when porosity passes across the interval from 30 to 50 percent. In this interval heat conduction of the soil drops from 0.0014 to 0.0004 cal/cm sec degree.

Figure 12. The effect of soil porosity P on its heat conduction λ (as per A. F. Shudnarskiy).

It should be pointed out that the porosity of virgin soils ranges within the limits of 30 to 40 percent, while the porosity of tilled soils may attain a value of 60 percent and over. Of still greater importance is the humidification of the soil, i.e. the substitution of low heat conduction air by water, the heat conduction of which is 20 times greater. The increase in the heat conduction of the soil with greater humidification is shown in Table 10.

TABLE 10

Heat conduction of sand with relation to its moisture content

Water content (in percent)	0	5	15	20
Heat conduction (in cal/cm sec degree)	0.0008	0.0011	0.0019	0.0025

The first additions of moisture, in particular, increase the heat conduction sharply. With subsequent additions of moisture, the increase in heat conduction becomes more gradual.

Since soil humidity is subject to annual and, in the top-soil layer, to diurnal variation, the heat conduction of the soil, too, is subject to diurnal and annual variation, decreasing with the desiccation of the soil and increasing with its humidification.

Corresponding for 0400 and 1600 hours characterizing the diurnal variation of soil heat conduction are cited in Table 11 (as per A. F. Chudnovskiy).

The heat conduction of the soil is determined under laboratory or field conditions, with the last method the more effective, since it takes into account all means of heat transfer, taking place at a given moment in the soil that is being examined. (See A. F. Chudnovskiy, "The physics of the heat exchange in the soil", 1948.)

TABLE 11

Date	Humidity of soil (in percent)		Heat conduction (in cal/cm sec degrees)	
	0400 hours	1600 hours	0400 hours	1600 hours
18/VI 1940	7.0	6.4	0.0031	0.0026
19/VI 1940	5.7	5.1	0.0028	0.0026
24/V 1941	11.0	1.2	0.0034	0.0029
29/V 1941	13.4	0.6	0.0033	0.0029
8/VI 1941	10.7	8.8	0.0030	0.0027

The increase in soil temperature is determined not only by the amount of inflowing heat, but also by its thermal capacity. Thermal capacity is the amount of heat in calories required for the heating of one gram (by weight) or one cubic centimeter (by volume) of a substance by one degree Centigrade.

It is necessary to differentiate between specific thermal capacity (by weight) and volumetric thermal capacity. In analyzing the thermal economy of the soil we will deal predominantly with volumetric thermal capacity.

Table 12 cites the thermal capacity of soil components, with the exception of water, the thermal capacity of which is taken as unity.

The volumetric thermal capacity of the soil, as a whole, also depends on its porosity and moisture content, it being the case that, in contradistinction to heat conduction, thermal capacity may be computed as the mean suspension of the thermal capacities of all the soil components: solid particles (mineral and organic), water, and air.

TABLE 12

Soil components	Specific (by weight) thermal capacity (in cal/gram degrees)	Volumetric thermal capacity (in cal/cm ³ degrees)
Sand and clay particles	0.18 - 0.23	0.49 - 0.58
Peat	0.48	0.60
Air	0.24	0.0003

To compute the thermal capacity of the soil, it is necessary to know the volumetric weight of the soil in its undisturbed structure, its moisture content (not in percent, but in grams), and the volumetric thermal capacity of the solid particles of the soil.

Such a computation is presented in Table 13.

TABLE 13

Soil components	Volume occupied by (in cm ³)	Volumetric thermal capacity of a volume unit (in cal/cm ³ degrees)	Volumetric thermal capacity of actual volume (in cal/cm ³ degrees)
Solid substratum	0.56	0.49	0.27
Water	0.31	1.00	0.31
Air	0.13	0.00	0.00
Thermal capacity of soil			0.58

Due to the fact that water, on the average, has a thermal capacity twice as great as the mineral part of the soil, the volumetric thermal capacity of the soil will greatly increase with the increase in its moisture content, it being the case that the greater the porosity, the greater the variations.

The values cited in Table 14 show that the thermal capacity of peat, as compared to other soils, is at its lowest when dry, and at its highest when damp. The last condition is stipulated directly by its high porosity.

TABLE 14

Volumetric thermal capacity of soils with various moisture contents (moisture content expressed in percent of maximum moisture capacity)
(in cal/cm³ degrees)

Type of soil	Moisture content (in percent)			
	0	20	50	100
Sand	0.35	0.40	0.48	0.63
Clay	0.26	0.36	0.54	0.90
Peat	0.20	0.32	0.56	0.94

Table 15 shows the thermal capacity of sub-argillaceous soil in a certain locality of the Leningrad area. It was computed by proceeding from the mean perennial moisture content of this soil, subject to the condition that the specific gravity of its solid phase is 2.8 grams/per cm^3 , its volumetric thermal capacity is 0.5 cal/cm^3 , the porosity of the arable horizon is 0.50 percent, and the porosity of the sub-surface horizon is 0.40 percent.

TABLE 15

Volumetric thermal capacity of the soil under
the conditions of the Leningrad Region
(in cal/cm^3 degrees)

Months	Depth (in centimeters)					
	0 - 3	20	30	50	100	0 - 100
I	0.79	0.77	0.62	0.46	0.46	0.56
VII	0.46	0.53	0.46	0.37	0.46	0.45

Proceeding from the data presented in Table 15, it is possible to compute the amount of heat required to warm a 100-centimeter column of soil of one cubic centimeter in cross section by one degree Centigrade. In January, it will take $0.56 \times 100 = 56$ calories, in July, the respective number of 45 calories, i.e., by 20

percent less. This variation is based exclusively on the dynamics of the moisture content in the soil. Hence, the importance of taking into account the moisture content of the soil in thermal capacity computations becomes clear.

The thermal capacity of plants also depends on the water content in their tissues.

Throughout the growth of a plant, the water content in its tissues is subject to variations, naturally resulting in variations of its thermal capacity. These considerable variations in the water content of plants are shown in Table 16.

TABLE 16

Moisture content in percentage of total weight
of the green mass (wheat)

Point of observation	Fascicular stage	Full inflorescence	Lactescent ripeness	Caraceous ripeness	Full ripeness
Temir	75	75	47	33	25
Omsk	73	70	63	38	32

An approximate computation of thermal capacity shows for the phase of full inflorescence of the plant in Temir a gravimetric thermal capacity of 0.9 cal/gram degrees, and for the phase of ripening -- 0.6 cal/gram degrees.

In characterizing the heat exchange in the soil the temperature conduction coefficient K which equals the heat conduction coefficient λ divided by the volumetric thermal capacity of the soil c_p is most frequently used. K is numerically equal to the rise in temperature, sustained by a unit of volume of the soil as a result of the inflow of an amount of heat that equals λ :

$$K = \frac{\lambda}{c_p} \quad (8)$$

The dimensionality of the coefficient of temperature conduction is expressed in cm^2/sec , but in order to avoid small values for K it is frequently expressed in cm^2/min .

The coefficient of temperature conduction is of particular interest, since it is specifically this coefficient that determines the temperature distribution in the soil by depth. In soils with a low coefficient of temperature conduction, the diurnal and annual ranges of temperature are extinguished at lower depths, as compared with soils of a higher temperature conduction.

The deep layers of a soil with a low temperature conduction are slow in sustaining a rise in temperature, but, on the other hand, they are also slow in cooling. The surface layers of a soil with low temperature conduction, on the contrary, are distinguished by greater temperature ranges, and warm and cool more rapidly.

The lag of temperature maxima and minima with depth is determined directly by the coefficient of temperature conduction.

The coefficient of temperature conduction depends on the moisture content of the soil, but this dependence is a complex one, since the moisture content increases not only the heat conduction, but also the thermal capacity of the soil. During the first stages of humidification, the increase in the heat conduction of the soil is more intensive than the increase in its thermal capacity, and temperature conduction increases. With a further increase in moisture, the increase in heat conduction becomes relatively slower, as a result of which temperature conduction is decreased.

Figure 13 is a graphic presentation of the temperature conduction variation of mildly leached out black earth, with relation to the moisture content of the soil (data presented by S. I. Kostin).

cm²/sec

Moisture content of soil

Figure 13. Variation of temperature conduction of soil in relation to its moisture content (as per S. I. Kostin)

The above graph shows that, under field conditions, a dry soil (moisture content 10 percent) has practically the same temperature conduction as an amply humidified soil (moisture content above 30 percent).

Toward the surface, in the layer from five centimeters deep to the top, temperature conduction usually drops sharply. This is most frequently due to the presence of sod, which, like peat, has a low temperature conduction. During dry summer days, the decrease in temperature conduction of the surface layer is determined by its complete shriveling.

According to computations by T. L. Golubovaya made at the Central Geophysical Observatory imeni A. I. Voyeykov, the mean perennial temperature conduction of the mineral-containing soils in the Leningrad Oblast, in the 20 to 160 centimeter layer is $0.003 \text{ cm}^2/\text{sec}$, but at some points it attains a value of $0.005 \text{ cm}^2/\text{sec}$ (Leningrad, Lyesnoy -- arenose soil, Parfinskaya School of Forestry -- sub-argillaceous soil). The temperature conduction of peat soil (suburban station Zamosh'ye) is $0.001 \text{ cm}^2/\text{sec}$.

Under the desert conditions of Central Asia, in loess soils, in the environs of the city of Azysya, the temperature conduction at the depth of 6 to 20 centimeters is $0.003 \text{ cm}^2/\text{sec}$, with a moisture content of 8 percent, while the temperature conduction of the surface layer, under the same conditions, was less than $0.001 \text{ cm}^2/\text{sec}$, with a moisture content of 2 percent.

Special devices exist for the determination of the temperature conduction of the soil, particularly a device designed by A. F. Chudnovskiy, but it can also be determined by observations of the temperature distribution in the soil. The D. L. Laykhtman precision method for the determination of the coefficient of temperature conduction is very complex. We will, therefore, cite the approximation method of computation by the diminution of the temperature range with depth.

The formula for the approximate computation of the coefficient of temperature conduction is as follows:

$$K = \frac{[0.43(z - z_0)]^2}{l_g A_0 - l_g A_z} \cdot t, \quad (9)$$

where z_0 and z are soil depths in centimeters, for which depths the diurnal and annual temperature ranges are known;

A_0 and A_z are the corresponding temperature ranges; t is a constant factor. When computing K by the diurnal temperature ranges of the soil, this constant factor equals 0.36×10^{-4} when computing by the annual temperature ranges, it equals

$$0.99 \times 10^{-7}$$

Let us cite an example of computing K by diurnal temperature range.

Given

$$z_0 = 5 \text{ cm}, \quad z = 20 \text{ cm}, \quad A_0 = 7.1^\circ, \quad A_z = 1.7^\circ$$

Substituting into the formula the difference of the depths and the difference of the logarithms of the temperature ranges, we derive

$$K = \left(\frac{0.43 \cdot 15}{0.85 - 0.23} \right)^2 \cdot 0.36 \cdot 10^{-4} = 0.0039 \text{ CM}^2/\text{sec.}$$

The deficiency of this method is that at its basis is the assumption of a sinusoidal temperature variation (diurnal or annual) and the invariability of the coefficient of temperature conduction in depth and in time. The more the temperature variation departs from the sine curve, the greater the error. It is, therefore, necessary for computation purposes, to take periods with undisturbed temperature variation. The above method is also unsuitable for computations by the annual temperature variation involving soil layers subject to winter freezing, since the liberation and the absorption of the latent heat of melting distort the temperature variation curve.

Parallel computations by the D. L. Laykhtman precision method showed that the results of the approximation method are completely satisfactory for the warm part of the year for the 5 to 20 centimeter depth.

The formula by A. F. Chudnovskiy cited below is characterized by a higher degree of precision, since it takes into account not only the first term of the periodic series, into which the diurnal (or annual) temperature variation can be expanded, as is

done in formula (7), but also its second term.

In order to compute the coefficient of temperature conduction by the A. F. Chudnovskiy method, it is necessary to make four observations in 24 hours at two depths (Z_1 and Z_2), the hours to be equidistant from each other, as for instance, at 0100, 0700, 1300, and 1900 hours.

The formula is as follows:

$$K = \frac{3.14 (Z_1 - Z_2)^2}{2(a + t_g \theta)^2}$$

where

$$\theta = \frac{(\theta_1 - \theta_2)(\theta'_1 - \theta'_4) - (\theta_2 - \theta_4)(\theta'_1 - \theta'_3)}{(\theta_1 - \theta_3)(\theta'_1 - \theta'_3) + (\theta_2 - \theta_4)(\theta'_2 - \theta'_4)}; \quad (10)$$

$\theta_1, \theta_2, \theta_3, \theta_4$ is the temperature of the soil at the indicated ~~indicated~~ four moments at depth Z_1 ; $\theta'_1, \theta'_2, \theta'_3, \theta'_4$ is the temperature of the soil at the indicated four moments at depth Z_2 ; T is the diurnal period, which is 86400 seconds.

Examples of the diurnal heat exchange variation on clear days are given in Table 17 (minus means that the heat is flowing from the active surface into the depth of the soil).

TABLE 17

Diurnal variation in the heat exchange in the soil
(cal/cm² min)

Points of observation	Observation period		Hours											
			0100	0300	0500	0700	0900	1100	1300	1500	1700	1900	2100	2300
Arys' (desert)	25 August to 12 Sep- tember 1945	45°	0.08	0.07	0.05	-0.06	-0.13	-0.17	-0.13	-0.05	0.05	0.08	0.07	0.08
Koltushi (meadow)	10 days in July 1948	60°	0.05	0.04	-0.01	-0.05	-0.10	-0.09	-0.07	-0.04	0.03	0.04	0.06	0.05

The two climatically extreme examples cited in Table 17 make it possible to establish the general regularities of the diurnal heat economies of the soil during clear summer days. The maximum thermal flow into the soil occurs until noon. The flow of heat into the soil stops considerably before the sun sets. In both cases, at 1700 hours, the soil begins to give up the heat stored during the preceding hours. The maximum outflow of heat occurs during the first half of the night. The value of the thermal current, in clear summer weather, is greater during the day than at night, yet even during the day, it does not attain $0.2 \text{ cal/cm}^2 \text{ min.}$

The soil cover (vegetation in the summer, snow in the winter) exerts a great effect upon the heat exchange in the soil. Data gathered by G. A. Lyuboslavskiy on the thermal balance of a soil, first denuded, then covered by grass during the summer, and by snow during the winter, from observations in Leningrad (Lesnoy).

The greatest diversity is observed in the winter from the moment of the establishment of the snow cover (December), which immediately stops the chilling of the soil, and in the spring, when the presence, first, of the snow cover, and then, the vegetation cover, retards the flow of heat into the soil. A similar effect is exerted by the vegetation cover upon the diurnal heat exchange in the soil, which of course, is reflected in its thermal cycle.

TABLE 18

Mean annual variation in the thermal balance of the
soil (from 0 to 160 centimeters deep) over
a period of 15 years, in calories

Soil	Months											
	XII- I	I- II	II- III	III- IV	IV- V	V- VI	VI- VII	VII- VIII	VIII- IX	IX- X	X- XI	XI- XII
Denuded	119	74	-79	-181	-511	-450	-175	111	300	249	250	257
Covered	48	40	16	-34	-421	-306	-170	14	212	220	288	120

Figure 14 shows the diurnal variation of the heat exchange in a denuded (black fallow) and in a grass-covered soil. It must be taken into account that the grass cover was low (not above five to seven centimeters in height), and the depicted variations are due to the effect of sod, which caused a sharp drop in the temperature conduction of the upper soil layer (in denuded soil, in the zero to five centimeter layer, $K = 0.0023$, and under the grass cover, in the same layer, $K = 0.0012 \text{ cm}^2/\text{sec}$).

cal/cm² min

B

A

Hours

Figure 14. Diurnal variation in the intensity of the heat exchange in denuded soil (A) and in grass covered soil (B). Observed at Koltushi, 8 to 9 July 1948.

A grass cover of great density can exert an even greater effect, it being the case that, in addition to a direct decrease in temperature conduction, the increase in the amount of the latent heat of vaporization also has its effect. A soil under a light grass cover without sod (for instance, a plowed parcel), on the contrary, can be distinguished by a heat exchange of higher intensity, since the diminution in the velocity of the wind in the midst of the grass cover reduces the heat exchange between the soil and the air.

The effect of wind velocity on the heat exchange in the soil, and, together with that, on its entire thermal cycle, was not adequately taken into account until recently. Yet, the functional relationship between the heat exchange in the soil and the velocity of the wind is an indispensable result of the unity in the thermal balance of the active surface. An increase in wind velocity induces a greater heat exchange between the soil and the air at the expense of the heat exchange in the soil itself.

The variation in the value of the nocturnal heat exchange in the soil, with relation to wind velocity, as per hourly observations during four clear nights (Leningrad, Koltushi), is represented in Figure 15.

cal/cm²min

meters/sec

Figure 15. Variation in the intensity of the heat exchange in the soil, with relation to wind velocity, observed during four clear nights in July and August 1948 (from 2000 hours to 0200 hours)

The velocity of the wind has a similar effect during the daylight hours.

Chapter 3

HEAT EXCHANGE BETWEEN THE SOIL AND THE AIR;
TURBULENT EXCHANGE; LATENT HEAT OF VAPORIZATION

The thermal flow from the soil into the air Q can be represented, similarly to the thermal flow into the soil, by the following approximation formula:

$$Q = c_p \rho K \frac{t_2 - t_1}{z_2 - z_1} \quad (11)$$

where c_p is the specific thermal capacity of the air, equal to 0.24 cal/gram degree; ρ is the density of the air at the terrestrial surface, equal to 0.0012 to 0.0013 gram/per cm³;

K is the coefficient of turbulent conductivity, otherwise known as the coefficient of turbulence, expressed in cm²/per second;

$\frac{t_2 - t_1}{z_2 - z_1}$ is the approximated vertical temperature gradient (provided the turbulence factor and the temperature gradient pertain to the same altitude).

The product $c_p \times \rho$ characterizes the volumetric thermal capacity of the air, equal at the surface layers to 0.0003 cal/cm³ degree, or 2000 times less than the volumetric thermal capacity of the soil.

The turbulence factor is of the same dimensionality and plays the same part in the heat transfer, taking place in the

air, as the coefficient of temperature conduction in the soil, which, considering its dimensionality ($\text{cm}^2/\text{per second}$) can be called the coefficient of conductivity. The coefficient of turbulence receives its name from the fact that, in the atmospheric air, the transfer of heat as well as other characteristics (humidity, dust content, quantity of motion) is not of a molecular nature as, for example, in solid bodies, but is carried with the aid of vortices (turbulentus -- vortex-like). The coefficient of turbulence is incomparably greater than the coefficient of temperature conduction of the soil and the molecular temperature conduction of the air ($0.163 \text{ cm}^2/\text{sec}$), and during one diurnal period, it can change tens, or even hundreds, of times. In the lowest atmospheric layer, it increases with height, in its first approximation, in direct ratio to the height. Molecular temperature conduction of the air is of practical value in the very lowest air layer only.

Let us dwell in some more detail on the mechanism of turbulent exchange.

Investigations on the variation of turbulence, with relation to weather conditions and peculiarities of location, indicate a leading part played by two factors: the dynamic and the thermal.

When there is a movement of air over a rough surface, vortices of various sizes are generated, in relation to the dimensions of the surface irregularities. We see these vortices

in the streets and in the fields when they carry with them dust, dry leaves, etc. As a result of the vortex motion, the air particles from the lower layers rise upward, and the ones from the upper layers descend. In addition, the generated vortex is carried by the air current in a fashion similar to the whirlpool funnels carried by a water current.

The vortex formation occurs at the expense of the kinetic energy of the wind and is increased with the increase in the latter. The greater the velocity of the wind, the more intensive the turbulent intermixing.

But, in addition to the dynamic energy source of turbulent intermixing, there is a thermal energy source, which not only induces the formation of vortices, but also has an effect on the displacement of all vortices in a vertical direction. In daylight during the warm part of the year, the unequal warming of the sunlit and the shaded side of a leaf, a stone, a hillock, and on a larger scale, black fallow parcel and meadow, body of water and forest, the sunny and shaded side of a street, induces the formation of vortices of a thermal origin. The superadiabatic temperature gradients of the air, usually observed at this time, promote the upward rise of the vortices, which were formed under the effect of thermal and dynamic factors.

As a result of the combined effect of the thermal and dynamic factors, turbulent exchange rises sharply in daylight, it being the case that with intensive "thermal convection"

(this is the name for turbulent exchange, when the thermal factor is predominant), the part played by the wind becomes secondary.

A substantially different situation occurs at night, with intensive radiation chilling and a temperature inversion caused by the latter. If vortices of a thermal origin develop during the night, they are considerably milder as a result of a smaller temperature differential, since the intensity of the nocturnal radiation loss of heat is much below that of the daylight hours heat inflow (see above). The most important fact is the stable stratification of the lowest atmospheric layers, which extinguishes any vertical displacement, with the exception of the downflow of the colder and therefore, relatively heavier air. The part played by the wind is particularly great during the night. The abating of the wind, and in connection with that, of the turbulent exchange, promote simultaneously the establishment of a temperature inversion, which, in its turn, determines the further abating of turbulence. Inversely, the intensification of the wind directly and indirectly increases turbulence by breaking atmospheric stability.

Extensive empirical data and theoretical considerations point to the fact that in open level areas, above the irregularities of the underlying surface (in particular, the vegetation cover), the turbulent intermixing in the lowest atmospheric layer increases with height in direct ratio to the height of the vegetation cover, provided the atmosphere is in a state of equilibrium,

i.e., the value of the temperature gradients is near zero. With superadiabatic gradients, turbulence increases with height more rapidly than in the case of a state of equilibrium, and in the presence of a temperature inversion, on the contrary, its increase with height is slower.

These characteristics in the variation of turbulence with height, in their turn, stipulate the vertical gradients of the wind, temperature, and humidity in the air. In the lowest 1-2 meter air layer over the active surface, the flow of heat and moisture can be considered as practically unvarying with height due to the low heat and moisture capacities of the air. As to the kinetic energy of the wind, i.e., the energy of the directional air current, it can be considered that the part of the latter consumed in the overcoming of friction and the formation of vortices, is consumed basically within the limits of the heights of the irregularities in the underlying surface. The kinetic energy current of the wind can also be considered as invariable within a layer several meters above the height of the active surface irregularities.

In the presence of an invariable current (see formula (11) for thermal current), the gradients are in an inverse ratio to the turbulence factor. And since, when there is a state of equilibrium, turbulence varies in direct ratio to height, the gradients, in this case, will be inversely proportional to height. In the presence of a superadiabatic stratification, the gradients

will vary not in an inverse ratio to height, but somewhat more rapidly, while in the presence of a temperature inversion -- somewhat more slowly.

The theory of turbulent exchange, as well as the computation methods for the determination of the turbulence factor, are still in their preliminary stages of development. Nevertheless, some methods may already be utilized in order to arrive at the approximated characteristics of turbulent exchange.

Below is cited one of the computation methods, offered by the Soviet scientist D. L. Laykhtman, and based on the follow-up of the variation of wind velocity with height. A method of greater precision could not be treated in these pages on account of its complexity. It is to be found in the thesis by D. L. Laykhtman and M. I. Timofeyev under the title "Turbulent exchange in the lower atmospheric layers", Transactions of GGO, Publication 20 (82), 1949.

The above mentioned simpler method was experimentally checked for an air layer ranging from 1 to 10 meters high. Due to the prior admission of the vertical invariability of the current which was taken as a basis of this method of computation, it can be utilized only for the air layer lying above the irregularities of the active surface. It cannot be applied in the case of an interlaced area within the range of heights of the irregularities. Nor can it be applied at the very surface of the soil, and in particular, in the midst of the vegetation cover.

According to D. L. Laykhtman, the coefficient of turbulence k meter²/second can be computed with the aid of the following formula:

$$K = \beta u_z z' - \alpha, \quad (11a)$$

where β is a variable magnitude, determinable by the variation in wind velocity with height, which in turn, is related to thermal stratification of the air and the roughness of the underlying surface; u_z is the wind velocity at the height of 2 meters; z' is the height, for which the coefficient of turbulent exchange is being computed; α is a magnitude related to the velocity of the wind, the thermal stratification of the air, and the roughness of the underlying surface, and also determinable by the variation of wind velocity with height.

Under roughness are understood the irregularities of the active surface, which increase the friction of the air current and decrease linearly its directional velocity, but at the same time increase the chaotic vorticity motion, or turbulence.

The method for determining α and the exponent of roughness z_0 along the vertical configuration of the wind is cited on printed pages 72 to 75. Figure 16 shows a graph of the curves,

with the aid of which, provided roughness z_0 and the value of α are known, it is possible to determine the value of β

When the atmosphere is in a state of equilibrium, the values of α approximate zero. In the presence of superadiabatic gradients, the value of α is negative, in the presence of a temperature inversion, it is positive.

In those cases, when α cannot be determined from direct observations on the vertical configuration of the wind, it is possible, for the approximated computations of K to utilize the observations on the velocity of the wind at the height of two meters and on the temperature differential in the air as between the 20-centimeter and the 150-centimeter layers.

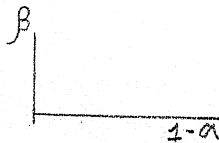


Figure 16. Nomogram for the computation of factor β in the formula for the coefficient of turbulence.

In Table 28 are cited the values of α with relation to wind velocity w_2 and thermal stratification of the air $\Delta t_{20,150}$. The value of z_0 can be determined approximately by the characteristics of the underlying surface (see printed page 72).

The coefficient of turbulence computed in this manner is usually expressed in meters²/per second.

Alongside of this method, there is the method of M. I. Budyko and others. It is regrettable that all of them, including the D. L. Laykhtman method, result only in approximated values for the coefficient of turbulence. These, however, are adequate in a number of cases for the solution of practical problems.

We now pass to the analysis of the diurnal variation in the coefficient of turbulence, as depicted in Table 19 below, it being the case that, in order to compare it with the coefficient of temperature conduction in the soil, it is expressed in cm²/per second. During summer clear weather, the coefficient of turbulence at the height of one meter at night is less than 100 cm²/second, while in daylight it is more than 1000 cm²/second. Considering that the mean value of the coefficient of temperature conduction in the soil is 0.003 cm²/second, we conclude that, already at the height of one meter, the turbulent temperature conduction of the air is 100,000 and even 1,000,000 times greater than the temperature conduction of the soil.

TABLE 19

Diurnal variation of the coefficients characterizing
the velocity of heat transfer in the air at the height
of one meter over a steppe (mean summer values)

Coefficients	Hours											
	0100	0300	0500	0700	0900	1100	1300	1500	1700	1900	2100	2300
$k \text{ cm}^2/\text{sec}$	80	80	150	470	860	1170	1250	1250	1250	860	230	150
$A = K \rho \text{ gram/}$ cm sec	0.1	0.1	0.2	0.6	1.1	1.5	1.6	1.6	1.6	1.1	0.3	0.2
$C_p \rho K \text{ cal/cm-}$ sec/degree	0.02	0.02	0.05	0.14	0.26	0.36	0.38	0.38	0.38	0.26	0.07	0.01

In place of the coefficient of turbulence K , there is frequent use of the coefficient of turbulent exchange A , which equals the product of K by the density of the air ρ . In the second line of Table 19, the diurnal variation of the coefficient of turbulent exchange A (frequently referred to as coefficient of exchange) is given. The coefficient of exchange A in grams/cm second is numerically equal to the quantity of air in grams passing through one square centimeter per second.

Finally, in the last line of the Table is given the product $C_p \rho K$ cal/cm-sec/degree. This magnitude can be called the coefficient of the turbulent heat conduction of the air. It also varies in time and with height as does the coefficient of turbulence.

The coefficient of turbulent heat conduction of the air, at the height of one meter at night, is tens of times greater than same for the heat conduction of the soil, and in daylight it exceeds the value of the latter by hundreds of times (see Tables 10 and 11).

As already mentioned above, in the lowest atmospheric layer, all indices of exchange (K , A and λ), in their first approximation, vary in ratio to height. Thus, already at the height of 10 centimeters, their values will be 10 times less, and at the height of one centimeter, 100 times less. At the very terrestrial surface, in a layer, the height of which is measured in millimeters and fractions thereof, turbulent exchange is extinguished, and heat transfer, as well as other characteristics, are predominantly effected in a molecular way, with the coefficient of molecular heat conduction of the air equal to 0.00005 cal/cm sec, i.e., 100 times less than that of the soil. However, in the presence of thermal convection, turbulent exchange, even though weakened, is present at the very edge of the active surface.

Due to the abating of turbulent exchange, at the very lowest air layer, particularly at night, in the presence of a deep tempera-

ture inversion, it becomes necessary to reckon with heat transfer by way of long-wave radiation, the significance of which, as yet, we cannot define.

The sharp diminution in turbulent heat exchange in nearing the active surface is the basic cause for the equally sharp increment in the gradients of temperature, humidity, and wind in the same direction. This also explains the abrupt temperature jump observed at the very edge of the active surface.

In Table 20 there is data on the relationship between the coefficient of turbulence at the height of one meter and weather conditions, as observed on a meadow in the environs of Leningrad by T. A. Ogneva.

[See Table 20 on next page]

The coefficient of turbulence in the above table is given not in cm^2/sec , but in $\text{meters}^2/\text{per sec}$, in order to avoid large figures.

As already indicated, the above method of computing the coefficient of turbulence holds true for an exposed level locality. In shielded localities, in the midst of vegetation, buildings, and interlaced conditions of terrestrial relief, we as yet, do not know how to determine turbulence. In first approximation, it can be accepted, that, under the above conditions, the relationship to the turbulence over an exposed level place is

TABLE 20

Dependence of the coefficient of turbulence (K meter²/sec)
at the height of one meter on the weather

Cloud formation (in ball-units)	Warm season						Cold season		
	Wind velocity at the height of one meter (in m/sec)								
	up to 1		1 - 3		4 - 6		up to 1	1 - 3	4 - 6
	day	night	day	night	day	night	during diurnal period		
0 - 7	0.20	0.005	0.20	0.02	0.25	0.05	0.005	0.02	0.05
8 - 10	0.10	0.01	0.15	0.05	0.20	0.10	0.005	0.03	0.10

proportional to the relationship of the wind velocities, i.e., if we assume the wind velocity over a glade at the height of one meter to be 1.5 times less than the wind velocity over an open field, then the turbulence at the same height over the glade will be approximately 1.5 times less.

The abating of turbulent exchange, due to the locality being shielded or to low wind velocities, results in that the total amount of heat entering the atmosphere is diminished, and a greater amount of heat remains for the soil. The intensification of the thermal flow into the soil, is determined in this case by the greater values of the temperature gradients in the soil itself, which are due to the high temperature of the soil surface. Yet, the temperature of the lowest air layers in such cases is somewhat up, since, due to the abated exchange, a relatively greater part of the heat entering the atmosphere is retained by the lowest air layers than during a more intensive exchange.

Similar conditions are observed during nocturnal cooling, with the only difference, that, in that case, the heat is flowing from the air into the soil. With abated turbulence, the thermal flow from the upper air layers is decreased, and there is a particularly sharp chilling below.

A good example of the effect of wind velocity on the heat exchange is shown in Figure 17, which is a graphic presentation of the diurnal variation of all the components of the thermal balance, the wind velocity, and the temperature for two contiguous

days, in which different wind velocities prevail.

	Direction of wind	meters/sec
grams/cm sec		
cal/cm ² min		mb
		ball-units
	cloud formation	
		hours
		days

Figure 17. Diurnal variation of the thermal balance components and other meteorological elements during calm and windy weather.

Arys', August 1945.

During the day 30 August, the temperature of the air t_v and the temperature of the soil t_p are higher than on 31 August. The heat exchange T in the soil is greater on the 30th, regardless of the fact that, on the 30th, the intensity of the radiation balance R and the heat exchange soil - air (Q) are lower. These peculiarities are due to the fact that, on the 30th, the wind velocity at the height of two meters is 2 meters/per second, while on the 31st it is 4 meters/per second. Hence we conclude that by changing the velocity of the wind, we will be in a position to regulate the thermal cycle of the surface layers of the air and soil.

The coefficient of turbulence also varies on a macro-scale. It differs in various climatic zones only in this case the decisive effect is exerted not by the velocity of the wind, but by the thermal stratification of the lowest atmospheric layers and by the presence of a stable snow cover, which sharply reduces the effect of roughness in the underlying surface.

In the south of the USSR, due to the predominance of clear weather and of the great intensity of solar radiation, superadiabatic gradients prevail during the day, inducing a sharp increase in turbulence. Toward the winter, the daylight turbulence is diminished, but is still exceeding the nocturnal turbulence, which is due to the absence of a stable snow cover. The annual cycle of nocturnal turbulence is of a reverse character. Clear summer nights induce temperature inversions leading to the diminution of turbulence. During the winter, the predominance of overcast weakens the temperature inversions and, consequently, increases turbulence.

In the northern and central parts of the USSR, the annual cycle of daylight turbulent exchange is also clearly manifested, but not at the expense of the summer maximum, which is smaller than in the south, but at the expense of the winter minimum. During the winter, due to the snow cover, the temperature inversion persists day and night, resulting in a low turbulence during the daylight as well as nocturnal hours. Due to the same cause, the maximum nocturnal turbulence occurs not in the winter, but with autumnal overcast and rains.

When the coefficient of turbulence and the temperature gradients are known, the heat exchange soil-air, including its diurnal variation, can be computed by formula (11).

Table 21 depicts the diurnal radiation of the heat exchange soil-air for the same two clear dry days of the warm part of the year at the same two observation points, for which the heat exchange in the soil was already cited (the minus sign means that the heat is flowing from the active surface into the air).

TABLE 21

Diurnal variation in heat exchange
soil-air (in cal/cm² min)

Points of observation	Hours											
	0100	0300	0500	0700	0900	1100	1300	1500	1700	1900	2100	2300
Arys'	0.01	0.01	0.01	-0.02	-0.15	-0.20	-0.20	-0.17	-0.07	0.01	0.03	0.02
Koltushi	0.01	0.01	0.00	-0.04	-0.12	-0.18	-0.20	-0.15	-0.08	0.00	0.01	0.01

The difference in intensity between the daylight and the nocturnal thermal flows is immediately evident. This phenomenon

is linked directly with the diurnal variation in the turbulent heat conduction as well as in the radiation balance itself (see above).

The second feature is that the thermal flow from the surface of the soil into the air begins only some time after sunrise and is discontinued somewhat ahead of sunset. The maximum thermal flow into the air is somewhat displaced during the postmeridian hours.

The coefficient of turbulent exchange A is utilized not only for the characteristic of the thermal flow into the air, but also for the computation of the value of vaporization from the active surface along the vertical gradient of specific humidity, provided that the vertical transfer of the water vapors is tied in either with vaporization or condensation (specific humidity is the quantity of water vapors expressed in grams per one gram of humidified air).

The greater the vertical differential in water vapor content (i.e., the greater the vertical gradient of specific humidity), the more of it will be carried from one level to the next one by one gram of air. In addition, the sum total of this transfer is tied in with the number of grams of air, passing through one square centimeter in a unit of time, i.e., with the coefficient of exchange A .

By using the approximated formula (12) below, it is possible, by the vertical gradient of specific humidity and by

by the coefficient of exchange, to compute the value of vaporization.

$$e = A \frac{f_2 - f_1}{z_2 - z_1} \quad (12)$$

where e is vaporization in grams/cm² sec, $\frac{f_2 - f_1}{z_2 - z_1}$ is the approximated value of the vertical gradient of specific humidity (provided, the coefficient of exchange and the gradient pertain to the same height).

If the right-hand member of the equality is multiplied by the value of the heat of vaporization d , we will derive the value of the loss of heat for vaporization E in cal/cm²sec:

$$E = Ad \frac{f_2 - f_1}{z_2 - z_1} \quad (13)$$

This formula is analogous with the formula for the heat exchange soil-air, if the latter is presented as follows:

$$Q = Ac_p \frac{t_2 - t_1}{z_2 - z_1} \quad (14)$$

by the coefficient of exchange, to compute the value of vaporization.

$$e = A \frac{f_2 - f_1}{z_2 - z_1} \quad (12)$$

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This formula is analogous with the formula for the heat exchange soil-air, if the latter is presented as follows:

$$Q = Ac_p \frac{t_2 - t_1}{z_2 - z_1} \quad (14)$$

The utilization of these formulas is somewhat inhibited by the difficulty in determining the coefficient of exchange, which is still further complicated by its great variability.

Nevertheless, such computations to derive the values of the heat exchange and vaporization are more and more coming into use. Most frequently, the diurnal variation of the above processes is characterized with the aid of these formulas. As is known, it is practically impossible to judge the diurnal variation of vaporization by the Rykachov or Popov vaporizers.

Over prolonged periods of time (such as a diurnal period, or a season), for which evaporation can be determined with the aid of vaporizers, or by the moisture balance in the soil, the heat exchange soil-air can be computed as the residual term of the thermal balance, by formula (1):

$$-Q = R + T + E .$$

In such determination of the heat exchange, the error will equal the sum of errors in the three remaining components of the thermal balance, which, as is known, are rather considerable. This is the basic defect of such computations.

The losses of heat to vaporization play a leading part in the thermal balance of the active surface. As is known, 600 calories of heat are required for the vaporization of one gram of water.

Table 22 depicts the mean values for the latent heat of vaporization, as per observations at the Novgorod station over the period 1922 to 1931.

TABLE 22

Latent heat of vaporization (in cal/cm² min)

Evaporating Surface	V	VI	VII	VIII	IX
Black fallow	123	141	176	111	59
Moss swamp	159	165	176	117	77
Winter rye	218	224	159	81	35
Common oats	153	230	212	100	47
Artificial meadow	181	212	240	123	77
Approximated daylight radiation balance	230	230	230	160	70

With relation to the character of the active surface, the latent heat of vaporization (including transpiration) varies substantially. The mean value of heat lost to vaporization over the vegetation period by a meadow is 30 to 40 percent

higher than same lost by a black fallow parcel, and winter rye in May absorbs 80 percent more calories for vaporization than the same black fallow parcel. The greatest losses of heat to vaporization occur in June and July. Toward September, the losses of heat to vaporization are sharply reduced.

The period of maximum vaporization is determined by a favorable combination of moisture reserves in the soil, a great evaporating surface of a well developed vegetation cover, and the presence of radiation heat. Towards autumn, the radiation balance is diminished, in connection with which vaporization is also reduced.

The process, which is the reverse of vaporization, is condensation; it is accompanied by the liberation of heat. The condensation of moisture on the active surface, in the form of dew or hoarfrost, is considerably lower in intensity than vaporization.

To characterize the diurnal variation in vaporization, we will use the data collected at the same two observation points: Arys' and Koltushi (Table 23) (the minus sign means that heat is lost to vaporization).

TABLE 23

Diurnal variation in thermal economy tied in with
vaporization-condensation (in cal/cm² min)

Observation Points	Hours											
	0100	0300	0500	0700	0900	1100	1300	1500	1700	1900	2100	2300
Arys'	--	--	--	-0.03	-0.06	-0.05	-0.07	-0.05	-0.03	--	--	--
Koltushi	0.01	0.01	0.00	-0.07	-0.17	-0.27	-0.27	-0.23	-0.15	-0.01	0.01	0.01

In contradistinction to heat exchange in the soil, the maximum of heat lost to vaporization is shifted to the second part of the diurnal period, which is linked to the corresponding shift of the temperature maximum of the evaporating surface. The intensity of heat liberation during condensation is 15 to 20 times below the intensity of the latent heat of vaporization.

As was to be expected, the latent heat of vaporization varies substantially with relation to macroclimatic conditions. Under the desert conditions of Arys', the losses of heat are 2 to 4 times lower than at a meadow surface in the forest zone.

Observations of the diurnal variation of vaporization and transpiration show that any kind of intensive vaporization

is possible only at the expense of radiation heat. With inadequate radiation heat, vaporization leads rapidly to the establishment of a temperature inversion, to the abating of turbulence, and to the saturation of the air layers, contiguous to the active surface, with water vapors. It is only in the presence of intensive advection of warm and dry air, accompanied by powerful winds (the foehns and dry winds of the southeast), that vaporization may occur outside of the direct tie-in with the radiation balance, including the nocturnal negative radiation balance.

CHAPTER 4

COMPARATIVE CHARACTERISTICS OF THE THERMAL BALANCE COMPONENTS
IN THEIR DIURNAL VARIATION. LOCAL ADVECTION AS THE
FACTOR DETERMINING THE PECULIARITIES
OF MICROCLIMATE

Let us compare the diurnal variation of the thermal balance components during the summer, developing as much as possible their mutual connection, as typical for two different climatic conditions. Figure 18 shows the curve graphs representing the diurnal variation for all four components of the thermal balance, by the observations in Arys' (under conditions of desert), and Table 24 depicts the correlations between the radiation balance and its components for Kolushi (meadow) and Arys'.

Cal/cm² min

Hours

Figure 18. Diurnal variation in the components of the thermal balance. Arys', 25 August - 12 September 1945.

TABLE 24
Correlations between the components of thermal
balance (regardless of sign)

HOURS

Components	2400-	0200-	0400-	0600-	0800-	1000-	1200-	1400-	1600-	1800-	2000	2200-
	0200	0400	0600	0800	1000	1200	1400	1600	1800	2000	2200	2400
Koltushin (meadow)												
T R	71	67	-	28	26	14	12	8	15	-	87	86
Q R	22	17	-	24	30	32	34	36	40	-	8	9
E R	7	16	-	48	44	54	54	56	75	-	5	5
Arys' (desert)												
T R	93	91	85	56	39	41	32	17	-	95	68	86
Q R	7	9	15	17	44	47	51	63	-	5	32	17
E R	-	-	-	27	17	12	17	20	-	-	-	-

Note: A dash in the columns above means either the absence of data (on vapor condensation at Arys'), or the non-significativeness of the correlations during the hours when the radiation balance approximates zero.

In calm clear weather the leading and of maximum intensity by day and by night, in the north and in the south, is the radiation balance. The intensity of the radiation balance in the desert is of the same order of magnitude as in the forest zone. The relatively low radiation balance of the active surface in the desert, as was already pointed out, is due, first, to the great reflective capacity (35 percent) of the light loess soil, denuded of vegetation, and, secondly, to the considerable effective radiation due to the high temperatures (above 50 degrees centigrade) of the surface of the soil. The intensity of the radiation balance of the irrigated vegetation covered parcels, under conditions of central Asia, is, of course greater than in the north.

In the forest zone during daylight hours, the radiation balance component second in magnitude is the loss of heat to vaporization (over 50 percent of the radiation balance). In the desert due to the absence of moisture, this component occupies the last place, while the second significant component, after direct radiation, is the thermal flow into the air (up to 60 percent). In the forest zone the thermal flow into the air is in the third place. The thermal flow into the soil at the meadow occupies the last place, and in the desert - the third place in the radiation balance.

Practically during all hours of the diurnal period the sum of relationships of all other components of the thermal balance to the radiation balance equals 100 percent, which attests the fact that they (Q, E, T) have the same algebraic sign. In daylight these flows have a negative sign (heat flows from the active surface into the soil, into the air, and is lost to vaporization). At night - they have a positive sign (heat flows to the active surface from the soil, from the air, and is liberated in the process of condensation). When the radiation balance approximates zero, which on clear summer nights occurs during the morning and evening hours, the sum of relationships does not equal 100 percent (if taken without regard to the algebraic sign). This means that the indicated 3 components have also different signs. For example, in Koltushi, at 1600-1800 hours, the flow of heat into the air, as well as vaporization, and, consequently, the loss of heat to vaporization, was still continuing and the soil was already giving off heat. It being the case that specifically the flow of heat from the soil was insuring relatively intensive vaporization, regardless of the decrease in the radiation balance.

From the above and similar data the conclusion can be made that the variations in the flow of heat into the air are stipulated not only by and, possibly, not to such an extent, by the radiation balance, as by the latent heat of vaporization.

In those cases when the latter is great, as for instance, in a field or meadow particularly after a rain or artificial irrigation, the direct flow of heat into the air is diminished at the expense of heat lost to vaporization. The latter heat also enters

the atmosphere but only in a latent state, being liberated only during the condensation of water vapor in the upper layers of the atmosphere.

Hence, the conclusion can be made that all other conditions being equal, the daylight temperatures of the lowest atmospheric layers increase with a decrease in vaporization. The hot weather of the sultry summers, as well as the torrid heat of the deserts, are stipulated not as much by the characteristics of the radiation balance, as by a sharp decrease in the latent heat of vaporization. When increasing the loss of heat to vaporization, as, for instance, by artificial irrigation, we thereby decrease the direct flow of heat into the air and into the soil, which condition, first of all, will reduce the temperature of the lowest air layers and the upper layers of the soil.

It also follows from this that under conditions of the north where there is a thermal deficiency in the lowest air layer, it is possible to ameliorate this condition by reducing vaporization with the aid of drainage, mulching, etc.

During the nocturnal hours the correlationship of the components is changed. Although the radiation balance still remains the leading factor, the second place is taken by the thermal flow from the soil (from 70 to 95 percent). The thermal flow from the air is then from 5 to 30 percent. The liberation of heat as a result of condensation hardly amounts to 10 percent.

We will now analyze the part played by the individual components in the occurrence of first autumn frosts.

The thermal cycle of the lowest air layers and also of the surface of the soil and vegetation during the nocturnal hours is determined not only by effective radiation, but also by the condition determining as to what extent this nocturnal loss of heat by the active surface is compensated for by the flow of heat from the soil. In addition, of great importance is the depth of the air layer, from which the heat flows to the active surface. In the presence of low turbulent heat conduction, the loss of heat from the air does not occur outside the range of the very lowest layers, resulting in their sharp chilling.

The complexity of the phenomena that stipulate nocturnal chilling must be taken into account in the problem of combatting first autumn frosts.

The prognosis of the radiation frost basically reduces itself to the prognosis of the thermal balance components: the radiation balance, which depends on the degree of cloud formation and temperature; air humidity; the thermal current from the air, which is tied in with the wind; the thermal current from the soil, determinable in the case of denuded soil, by its dampness and compactness, and, in the case of vegetation covered soil, primarily by the capacity of the latter, and the possible liberation of the latent heat of vaporization.

In evaluating the empirical methods for the prognosis of the first autumn frosts, it is first of all necessary to analyze the congruence of the above cited physical essence of the phenomenon. In doing so, it is necessary to bear in mind that

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In evaluating the empirical methods for the prognosis of the first autumn frosts, it is first of all necessary to analyze the congruence of the above cited physical essence of the phenomenon. In doing so, it is necessary to bear in mind that

such a prognosis is applicable only to level localities. In the presence of interlaced terrestrial relief, an important part is played by the run-off of cold air, or the so-called local advection, to the analysis of which we now proceed.

In large level land parcels which are homogeneous by the character of the soil and vegetation, a decisive part in the formation of local climate and microclimate is played by the local radiation balance and the local characteristics of the active surface. The microclimate and the local climate of such land parcels we qualify as independent.

In those cases when small parcels of land differ by the characteristics of their active surface, from the surrounding area, as for instance, the slopes of a hill or ravine, an irrigated parcel in the midst of a non-irrigated area, their local climate and microclimate are determined not only by the characteristics of the active surface on the spot, but also by the transfer of air from over adjacent land parcels by local advection. Such microclimates and local climates, which manifest the predominating effect of local advection, we qualify as non-independent.

The presence of a horizontal transfer of heat and moisture effects a change in their vertical transfer. For instance, at an irrigated land parcel, the thermal flow even on a summer day may be directed from the air to the soil as a result of the damp soil and growing vegetation losing so much heat to vaporization and transpiration, that the temperature of the surface of the soil and vegetation becomes lower than the temperature of the air that is

being transferred from the surrounding non-irrigated area.

The division of microclimates into independent and non-independent is important, and is to be taken into consideration in all cases when combatting destructive meteorological phenomena.

The part played by local advection in the formation of microclimate and local climate is particularly great during the night, when, due to radiation chilling, a layer of colder and, consequently, heavier air is formed at the terrestrial surface.

The stable temperature stratification weakens the turbulent exchange. As a result of this, the air cooled at the terrestrial surface does not mix well with the upper air that has not cooled yet. In a locality with interlaced relief the heavy cold air flows off into the low spots. The flow-off of the cold air occurs from the slopes of hills, ravines, along valleys, tops of trees and bushes. The place of the flowing off cold air is taken by the as yet non-cooled air of the upper layers. As a result, lower nocturnal air temperatures occur not over land parcels with a maximum radiation chilling and a weakened flow of heat from the soil, but in places, into which the cooled air is flowing off.

Thus, low-lying places turn out to be more frost-fissured than high spots regardless of the fact that the damper and, consequently, better heat-conducting soils of the first promote the inflow of heat from the soil, which inflow compensates for the relatively greater share of the radiation heat loss, as compared with the dry soils of the more elevated places.

It is not correct to liken the flow-off of cold air to the flow-off of water. The flow-off of water is determined by the difference in the specific gravity of water and air, and, as is known, water is almost 1000 times heavier than air. The proportion between the weights of warm and cold air, with a temperature differential of 3 degrees Centigrade is only 1:1.01. Therefore, the flow-off will be effected only in the presence of a slope steepness in excess of 2-4 degrees. Vegetation, arboreal and undergrowth usually retards the flow-off of cold air. Figure 19 shows a distribution diagram of the nocturnal temperature under condition of interlaced relief resulting from the cold air flow-off.

Warm

Cold

Figure 19. Diagram of a nocturnal air temperature distribution under condition of interlaced terrestrial relief.

The capacity of the flow-off air current is determined by the air collecting area. On the top of a slope it is small - less than one meter. At valley bottoms it may attain a value of tens of meters and more and it may be transformed into commonly known mountain breezes. The upper boundary of such a current is clearly outlined. The flow-off velocity is most frequently less than

1 meter/ per second. The flow-off of cold air and the temperature differentiation tied in with it are induced by an intensive radiation chilling and a mild turbulent intermixing; in other words, by clear calm weather. We shall return to the subject of the cold air run-off in the section devoted to the effect of terrestrial relief upon the frost hazards of a territory.

SECTION II

THE BASIC CHARACTERISTICS OF THE VERTICAL STRATIFICATION OF TEMPERATURE AND WIND IN THE LOWEST ATMOSPHERIC LAYER

Chapter 5

Variation in Temperature with Height in the Lowest Atmospheric Layer - Microoscillations of Temperature.

The vertical temperature variation in the lowest atmospheric layer is tied in directly with the heat exchange between the soil and the air.

During the day when the heat flows from the surface of the soil into the air the temperature diminishes with height. At night, with the flow of heat in the opposite direction, temperature rises with height, i.e., a temperature inversion.

Table 25 depicts an example of the temperature variation with height over black fallow, according to observations in the environs of Leningrad.

TABLE 25

Temperature of the air over denuded soil,

Sablino, 15 July 1938

Hours	Height in Cm				Height differentials		
	5	25	50	150	5-150	5-50	50-150
	Temperature				Temperature differentials		
1200	31.6	30.5	29.7	29.2	2.4	1.9	0.5
2100	19.5	20.3	20.7	21.3	-1.8	-1.2	-0.6

It is notable that the basic variation occurs in the very lowest air layer.

The decrease in the temperature gradient with height may be induced by two causes: (1) diminution of the thermal flow with height and (2) an increase in turbulence since in accordance with the thermal current formula (11), the temperature gradient is in direct ratio to the current and in inverse ratio to the coefficient of turbulence. The first circumstance, under steady conditions, has no significance for the temperature gradients in the lowest layer, since, due to the small volumetric thermal capacity of the air, the thermal current practically does not vary in passing the lowest air layer. The second circumstance plays the decisive part, since the coefficient of turbulence being, as indicated above, infinitesimally small at the earth's surface, increases with height approximately in proportion to the latter (see Chapter 3). If the temperature gradients were decreasing in an inverse ratio to height, the temperature itself was diminishing (during the day) or

increasing (during the night) in proportion to the logarithm of height, since a derivative of specifically a logarithmic function (the vertical temperature gradient is a derivative of temperature by height), as is known, is inversely proportional to the magnitude itself.

In this case, the vertical configuration of temperature [plotted] by semi-logarithmic coordinates (the logarithms of heights are laid off along the ordinate, the temperatures - along the abscissa) would be represented by a straight line. But since the variations in gradients by height are somewhat parted from the indicated regularity, the variation of the temperature itself with height does not fully conform to a logarithmic variation, by the virtue of which circumstance its configuration [plotted] by semi-logarithmic coordinates shows some curvature.

Figures 20 and 21 represent the same configuration of temperature plotted in different coordinate [systems]: in figure 20 the temperature profile is plotted by coordinates height-temperature, in Figure 21 - by coordinates logarithm of height - temperature. The observations were made in the environs of Leningrad, in an exposed locality, with the aid of a ventilation psychrometer. For the observations at the higher points, a triangulation tower was used.

Height

Temperature deflection at the height of 1.5 meters

Figure 20. Variation of temperature with height, in temperature deflections, at the height of 150 centimeters, as observed in clear calm weather, at Vsevolozhskaya. July - September 1945 (as per N. V. Smirnova).

Designation of heights: 1-5, 2-20, 3-50, and 4-150 centimeters.

Cm by Cm

Height

Temperature deflections at the height of 1.5 meters

Figure 21. Variation of temperature with height (same as in Figure 20, except plotted by semi-logarithmic coordinates).

Figure 20 makes it possible to establish the basic characteristics of the daylight and nocturnal configurations of temperature. The temperature drop with height during the day and its rise with height during the night are in themselves not fully characteristic of the phenomenon. Although in both cases the greatest variations occur in the very lowest layer, the daylight variation in the layer up to 1.5 meters exceeds the variation in the 1.5-34 meters layer,

while, in the nocturnal variation, conversely, the temperature rise in the 1.5-34 meters layer is greater than the temperature rise in the very bottom air layer.

The fact that at night the temperature gradients at the very surface of the soil are smaller than during the day is due to the circumstance that the source of the nocturnal flow of heat - the radiation balance (see printed page 31) is smaller in value than during the day. The greater value of the nocturnal temperature gradients, in the air layer above 1.5 meters, is determined by their slower decrease with height, which, in turn is due to a slower (than during the day) increase in turbulence.

Plotted by semi-logarithmic coordinates, the daylight configuration of temperature is almost a straight line. This indicates that temperature diminution during the daylight hours is close to logarithmic in character. But if the ruler is applied to the two upper points, some curvature, with its convexity toward the abscissa, can be discerned. The nocturnal temperature configuration, however, shows a plainly evident curvature. Up to the height of 1.5 meters, the temperature differential of 0.2 degree Centigrade corresponds to the difference in the logarithms of heights, which is approximately 0.5. At the height of 10-20 meters, a temperature differential of 0.4 degree Centigrade corresponds to the respective difference in the logarithms of heights.

Since the temperature gradients are dependent on the source of heat and turbulence, they can vary quantitatively to a substantial degree in relation to weather conditions and the character of the

underlying surface, but their indicated characteristics are usually preserved. Only an increase in wind velocity, independently of other factors, causes the vertical configuration of temperature to approximate the logarithmic. The relative proximity to the logarithmic configuration, particularly in daylight, must be taken into account in the solution of practical and theoretical problems, including the interpolation of observation data, and, in some cases, even in extrapolation. This, specifically, is the reason for plotting the vertical configuration of temperature in semi-logarithmic coordinates, since interpolation and, particularly, extrapolation, is much simpler done along straight lines. The semi-logarithmic coordinates are convenient also in that they make it possible to incorporate into the graph the temperature in a layer, from several centimeters to tens and even hundreds of meters, with detail, corresponding to the temperature variation, without necessitating an excessively large drawing.

It is also necessary to take into account the near-logarithmic vertical distribution of temperature in the selection of heights for observations. The distances between observation points must be in ratio to the logarithms of the heights, i.e., they must increase with height. The indicated characteristics in the distribution of temperature are preserved up to a height of several tens of meters.

During the transition hours of the diurnal period, and also with a sharp change in the wind or in the intensity of the radiation balance, the vertical temperature configuration becomes deformed. It may happen that, during the evening, there is already a temperature inversion below, while in the higher layer a weak

superadiabatic gradient still persists. In the presence of a ground radiation fog, the inversion gradient may attain a second maximum value at the upper boundary of the fog.

Under conditions of interlaced relief, and also in areas shielded by vegetation or structures, the vertical configuration of temperature undergoes substantial deformation, in conformity with turbulence deformation. Within the limits of the height of the shielded zone, the gradients decrease, after which there is an abrupt upward jump. The configuration of temperature in the midst of a grass cover and in a forest will be analyzed in the respective chapters.

The nature of the thermal stratification of the lowest atmospheric layers is of great practical importance, particularly due to the fact that it controls turbulent exchange and all the processes tied in with the latter, first of all as important a process as vaporization.

During a temperature inversion, natural ventilation is weakened, which circumstance is to be taken into account in industrial centers, where great quantities of waste gases and smoke enter the atmosphere.

Of greatest practical importance is the determination of the time the temperature inversion begins, and the time it ends, since such a determination segregates the period of weakened turbulent exchange.

Corresponding research into the thermal balance of the active surface, particularly the observation in Arys' and in the environs of Leningrad, confirm the parallelism in the course of the radiation balance and the course of the heat exchange between the soil and the air, which circumstance can be utilized for the computation of the diurnal variation of the temperature gradients.

Of the components of the thermal balance (the thermal economy of the soil and the heat exchange linked to vaporization and condensation), the heat exchange soil-air, by its general diurnal variation (but not by its absolute value and with a reverse sign) approximates the closest to the diurnal variation of the radiation balance (see Figure 18). But the moments of their passing through zero value do not fully coincide. The positive thermal flow into the air (and, consequently, the positive value of the temperature gradient) in the morning is lagging somewhat behind the corresponding value of the radiation balance, and in the evening it is discontinued before the latter. In this relationship, the leading part is played by the latent heat of vaporization. In the evening when, due to low relative humidity, vaporization is particularly intensive, the divergence of the above indicated corresponding moments is in excess of 1.5 hours, while in the morning it is equal to one hour.

The decisive role of vaporization is confirmed by parallel observations of the temperature gradients over a wheat field and over a black fallow soil area, which were conducted in 1939 and 1940 by the All-Union Institute for Plant Cultivation. Over the

black fallow soil area, the total vaporization from which is considerably below the total vaporization from the wheat field, the negative temperature gradient (the temperature inversion) usually manifested itself one hour behind the same over the wheat field.

It can generally be accepted that the lower the value of vaporization, the closer the moments of the zero values of the radiation balance and the heat exchange between the soil and the air approximate the point of coincidence.

Figure 22 shows a graphic comparison of the time of the setting in and the discontinuation of the temperature inversion with the corresponding solar altitudes, according to the observations conducted by Best in Salisbury (South Anglia).

The graph shows that the mean monthly zero value of the temperature gradient Δt is timed to the solar altitude of 10-15 degrees.

hours

Figure 22. Time of the setting in and discontinuation of the nocturnal temperature inversion in the air layer 50-120 centimeters, as observed by Best (= 51 degrees Centigrade), in 1. $\Delta t=0$; 2- $h_0 = 10$ degrees Centigrade; 3 - $h_0 = 15$ degrees Centigrade.

Figure 23 depicts similar data obtained by the expedition of the Central Geophysical Observatory. Here, too, the zero gradients are timed to the solar altitude of 10 - 15 degrees, with a departure of not over 0.5 of an hour. The only exception is manifested by winter observations, in the presence of the snow cover, when the moments of the setting in and the discontinuation of the temperature inversion were timed to the solar altitude of 20 - 25 degrees. This confirms thereby the already mentioned above assertion about the effect of the great reflecting capacity (albedo) of the snow cover on the reduction of the period with positive radiation balance.

(Snow Cover)

hours

Figure 23. The time of the setting in and the discontinuation of the nocturnal temperature inversion in the air layer 20 - 150 centimeters, in relation to solar altitude (as observed by the expedition of Central Geophysical Observatory).

1 - $\Delta t = 0$; 2 - $h_{\odot} = 15^{\circ}$; 3 - $h_{\odot} = 10^{\circ}$

To determine the setting in and discontinuation time for the temperature inversion on individual days, see results of observations, shown in Figure 24, on the passing of Δt through its zero value at one of the observation points in the steppe zone of the USSR (observations were made over a period of several

months, morning and evening, every 15 minutes).

hours

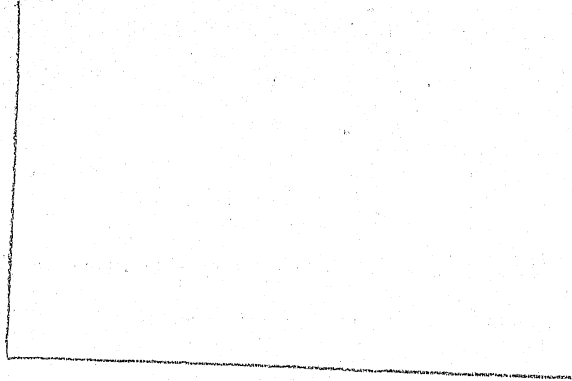


Figure 24. The time of the discontinuation (A) and the setting in (B) of the nocturnal temperature inversion in the air layer 20 - 150 centimeters ($\varphi = 55^\circ$), in relation to solar altitude.

As seen from the above, the departure of individual points from the 10 - 15 degree solar altitude (vertical line means that $\Delta t = 0$ was observed over the extent of the corresponding time) basically does not exceed ± 1 hour.

Almost all cases of an early setting in of the inversion are linked to intensive vaporization, which is confirmed indirectly by the moisture marks left on the surface of the soil by preceding precipitation. The cases of time lags in the setting in of the inversion are less significant, but on certain days the inversion was setting in only with the setting of the sun.

The moment of discontinuation of the inversion is even more clearly tied in with the solar altitude. The time lag in

the discontinuation of the inversion is basically linked to the degree of cloud formation. The maximum lag was observed in those cases, when a clear night was followed by an overcast morning. only on rare occasions was a considerable time lag in the discontinuation of the inversion observed under clear skies. This was due to precipitation that occurred the day before. Abundant dew is also conducive to the lengthening of the temperature inversion period at the expense of the increase in the latent heat of vaporization.

The vegetation of the steppe, as is known, transpires most intensively at the beginning of the summer, when, in the presence of high transpirational capacity of the young foliage and of great moisture reserves in the soil, the general transpiring green mass is adequately great. Toward the end of the summer, the steppe becomes dry, and total vaporization is diminished. Therefore, the temperature inversion period may be relatively longer at the beginning of the summer, primarily at the expense of the earlier setting in of the inversion. Yet, in first approximation, a solar altitude of 10 - 15 degrees, as the index of the zero value of the radiation balance (see Table 8), may still be utilized for the forecasting of the time of setting in and of discontinuation of the nocturnal temperature inversion in the 1.5 - 2.0 meter air layer with the mean precision of ± 1.0 hour.

Uniform method observations made at 10 locations in the USSR in 1943 - 1944, in exposed meadow grass areas, make it possible to determine the value of the midday (Δt_{pd}) and midnight (Δt_{pn}) temperature gradients in the air layer 20 - 150 centimeters in

their annual variation. As a conditional exponent of the gradient, we utilize the temperature differentials within the indicated layer. In conjunction with the time of the passing of the gradient through its zero value, they also characterize the diurnal variation of Δt . The effect of the weather was calculated separately for the daylight period and for the nocturnal hours. To determine the part played by solar altitude, mean values of t were taken from observations under clear (sky cover 0 - 2) and calm (wind velocity 0 - 2 meters/per second) weather, and with a dry surface of the soil.

Figure 25 shows the variation in the midday values of Δt in clear and calm weather, with relation to the solar altitude. It cannot be considered that, in this case, we are dealing with the direct effect of the radiation factor only. Alongside of the increase in solar altitude, the dryness of the soil is also increased. This phenomenon prevails under the climatic conditions of the USSR latitudinally (with the exception of the humid subtropics), as well as along the path of its annual variation (with the exception of "DVK"). This may cause an increase in the importance of the solar altitude factor as a complex exponent, since a diminution in the moisture content of the soil will induce a decrease in its thermal economy, and also in the latent heat of vaporization, increasing thereby the heat exchange between the soil and the air, and, consequently, of Δt_{pd} .

Figure 25. The relation of the midday Δt to solar altitude h 

1 - Anaseuli; 2 - Yemtsa; 3 - Yoshkar-Ola; 4 - Vysokaya Dubrava;
5 - Kuybyshev; 6 - Kyusyr; 7 - Mazanovo ("DVK"); 8 - Pyshkino-

Troitskoye; 9 - Repetsk; 10 - Tashkent.

Regardless of the fact that the above graph depicts localities as far apart as Kyusyur (eastern tundra), Tashkent, Yemtsa (Arkhangel'sk Oblast), Δt_{pd} is of the same order of magnitude and, secondly, is in adequately clear relation to solar altitude. This latter relationship may be represented by the following very simple formula:

$$\Delta t_{pd} = 0.028 h_0 - 0.42 \quad (14a)$$

Utilizing the above formula (14a), we will attempt to compute the annual variation of Δt_{pd} by latitudes. The results of such a computation are presented graphically in Figure 26. For the period of the stable snow cover (the cross-hatched part of the graph), the values of Δt_{pd} are not given, since, at that time of the year, the importance of solar altitude recedes into the background, and formula (14a) is not applicable.

Figure 26. The isopleths of the annual variation of Δt_{pd} by latitudes (excepting the period of stable snow cover, which is cross-hatched).

The values of Δt_{pd} obtained in such a manner correspond to clear and calm weather in exposed level areas covered with grass to a height not exceeding 10 centimeters. In the presence of other characteristics in the locality and in the soil cover, Δt_{pd} values will, of course, change (as is discussed in Chapters 8 and

9), as, for instance, the temperature of the soil changes under the effect of the vegetation cover. But, just as the soil temperature observations, conducted at meteorological stations under the so-called "natural cover", have a certain practical value, by the same token, the above cited values of Δt_{ps} may be utilized for the solution of certain problems, as, for instance, to bring out the regularities in the variation of the thermal stratification of the 20 - 150 centimeter air layer in macro-scale and to the extent to which they are stipulated by the latitudinal characteristics of the radiation balance and the moisture economy. The latitudinal correlations of the Δt_{ps} values must be generally preserved with relation to thermal stratification also over other active surfaces.

The midnight temperature gradients Δt_{pn} are, naturally, not related to solar altitude, but are determined by the degree of cloud formation, wind velocity, and the degree of soil dampness. The effect of soil moistening is confirmed by data presented in Table 26. The midnight Δt on clear and calm nights becomes greater when passing from damp to dryer zones, and there is also a tendency for the Δt_{pn} to become greater during the dryer summer season. It is of interest that the Δt_{pn} over the snow cover is of the same order as over the grass cover. Proceeding from the thermal characteristics of snow, its small degree of heat conduction, it could have been expected that the value of Δt_{pn} will be increased during the winter. This peculiarity is explained by the fact that effective radiation, under clear skies during the winter, is lower in value than during the summer,

which, in turn, is determined by the greater depth of the temperature inversion layer under winter conditions.

The damping of the active surface (after a dew or a rain) affects substantially the midday values of Δt

An increase in the latent heat of vaporization from a dampened surface reduces the direct flow of heat into the air, thereby reducing the daylight temperature gradient. Figure 27 is a graphic representation of the inverse ratio between the temperature differential in the 1 - 150 centimeter layer and the corresponding absolute humidity differential, in accordance with observations of the Maykop Black Fallow Experimental Station. In this case, the absolute humidity gradient can be considered as the index of the intensity of vaporization. High temperature gradients occur only in the presence of low humidity gradients. Inversely, when the humidity gradients are high, the temperature gradients become sharply diminished. In the mean, during the day, the value of

Δt_{pd} over a dampened surface is 60 percent of the value of Δt_{pd} over a dry surface. During the nocturnal hours, this relationship could not be established. The indicated diversity between the daylight and the nocturnal gradients may be explained by the fact that the dampening of the surface proper merely has a substantial effect on vaporization, which, at night, is generally discontinued and has little effect upon the thermal properties of the soil, such as the heat exchange within the soil, which is the leading factor during the night.

TABLE 26

Mean values of the midnight Δt in clear and calm
weather by climatic zones (the number of
observations indicated parenthetically)

Climatic Zones	Winter	Spring	Summer	Autumn
Damp subtropics (Anaseuli)	-0.6(35) ¹	-0.3(31)	-0.4(11)	-0.3(33)
Forest zone (Yemtsa, Yoshkar-Ola, Vysokaya Dubrava)	-0.5(87)	-0.6(39)	-0.5(98)	-0.3(41)
Far East forest zone (Mazanovo)	-0.8(25)	-1.1(19)	-0.7(25)	-0.6(26)
Steppe zone (Kuybyshev) .	-0.9(10)	-0.6(6)	-0.8(25)	-0.5(15)
Desert zone (Tashkent Repetsk)	-0.6(93) ¹	-0.9(64)	-1.2(87)	-1.2(54)

¹Without snow cover

Δ temperature

Δ Absolute humidity

Figure 27. The inverse ratio between the vertical variations in temperature and absolute humidity of the air layers closest to the soil (1 - 150 centimeters) over black fallow soil at 1300 - 1400 hours (under a clear sky) -- as observed by the Maykop Experimental Station in May 1939.

The effect of cloud formation and wind velocity on the midday and midnight t values is shown in Figure 28. The isopleths indicate as to what percent of t_{maximum} under a clear sky and with a wind velocity less than 2 meters/per second, is constituted by the value of t under the respective weather conditions.

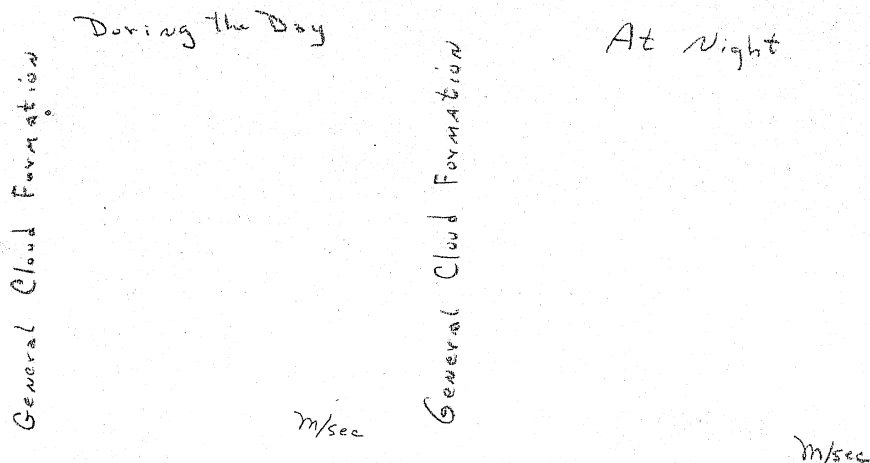


Figure 28. The variation in Δt in relation to cloud formation and wind velocity at a height of 2 meters, expressed in ratios of Δt , in the presence of 0 - 2 cloudiness and a wind velocity < 2 meters per second (in percent).

It must be noted that the extent of the effect of weather conditions varies with the time of day. The sharpest effect of cloudiness and wind velocity occurs at night. On an overcast night (with cloudiness rated at 8 - 10 balls), without reference to the wind, the value of Δt is less than 25 percent of

$\Delta t_{\text{maximum}}$. In the presence of a wind with a velocity of 5 meters per second or more, the value of Δt is less than 50 percent of $\Delta t_{\text{maximum}}$.

The peculiar characteristics of the above effect of cloudiness is due to the corresponding effect of the latter on the radiation balance of the active surface. During the day, the maximum radiation balance occurs not when the sky is perfectly clear, but rather in the presence of some small amount of cumuliiform cloud formation, when, alongside the almost continuous current of direct solar radiation, the diffused radiation

substantially increases, and the effective radiation becomes weaker.

During the night, the presence of even a low degree of cloudiness reduces effective radiation.

As to the effect of wind velocity, the data of the Arys' expedition of the Central Geophysical Observatory confirms the absence of a clearly pronounced link between the velocity of the wind and the daylight value of Δt . This is due to the fact that, during the day, the turbulent exchange, which determines the value of Δt , basically depends not on the velocity of the wind, but on thermal convection.

During the night, the velocity of the wind is decisive in the formation of turbulent exchange, thereby determining the clearly pronounced link of the nocturnal Δt with wind velocity, particularly when the latter is less than 2 meter/per second, as can be seen from Figure 29.

U_2 meters/second

Figure 29. Variation of Δt in relation to wind velocity at night (under a clear sky). Arys', 1945.

High vertical temperature gradients induce high micro-oscillations of temperature in time. Figure 30 shows results of observations by A. A. Znamenskiy and M. I. Gol'tsman with the aid of a low-lag platinum resistance thermometer and a ventilation psychrometer in Tashkent, the readings having been taken at two altitudes (5 and 100 centimeters).

Seconds

Figure 30. Microoscillations of the air temperature by psychrometer and resistance thermometer (as per M. I. Gol'tsman).

- A - Tashkent, h = 5 centimeters, 14 September 1933, 1210 hours;
- B - Tashkent, h = 100 centimeters, 10 September 1933, 1138 hours.
- 1 - resistance thermometer; 2 - ventilation psychrometer

Readings were taken every 5 seconds during near-midday hours, i.e. during the maximum development of thermal convection. Over a period of $2\frac{1}{2}$ minutes, the low-lag thermometer, at an altitude of 5 centimeters, registered temperature oscillations within the range 30.5 - 37.6 degrees Centigrade, the oscillation amplitude being 7.1. A considerably higher-lag instrument -- the ventilation psychrometer, registered oscillations within a range of 33.6 - 35.3

degrees Centigrade. At an altitude of 100 centimeters, during the same $2\frac{1}{2}$ minutes, the resistance thermometer showed an oscillation amplitude of 4.5 degrees, and the ventilation psychrometer -- an amplitude of 1.8 degrees Centigrade.

There are two types of temperature oscillations: minute ones and larger ones (gradual), as confirmed by other scientists. This is of great theoretical interest, since it is revelatory of the structure of the lowest air layers.

Research into the microoscillations in the lowest air layer led A. A. Skvortsov to the creation of the theory of the multi-stage stratification of convective exchange. He established that the temperature of the air on a clear summer day sustains oscillations of two types: (1) sharp oscillations with an average periodicity of 0.64 - 0.86 seconds, with the mean amplitude of the largest ones dropping with height from 1.5 degrees Centigrade at 5 centimeters to 0.6 degrees Centigrade at 400 centimeters; (2) regular slow oscillations (2 - 3 per minute) with a mean amplitude from 5 - 6 degrees Centigrade at the height of 5 centimeters to 1.5 degrees Centigrade at a height of 400 centimeters. According to A. A. Skvortsov, the process of convective exchange proceeds with the formation of multi-stage stratification, with closed circulation in each stage, and is accomplished by the periodic destruction and restitution of these stages. He considers as an established fact the presence in the lowest atmospheric layer of three such stages: the near-the-soil stage 3 - 4 centimeters deep, the lower near-the-terrestrial surface stage 2 - 3 millimeters deep, and the upper near-the-terrestrial surface stage up to

several hundred meters in depth. The indicated microoscillations are seemingly observed only in the presence of thermal convection. At night, in the presence of temperature inversions, the microoscillations are insignificant, not exceeding 1 degree Centigrade. Such insignificant temperature oscillations also prevail on days with overcast formed by low hanging clouds.

The temperature microoscillations are of great practical interest, and they must be taken into account in the evaluation of the thermal cycle of the environment for low-inertia bodies. They must also be taken into account in determining the methods of observation, since, due to their presence, a one-time reading of a low-lag thermometer may give a random non-characteristic value. Therefore, with the exception of special temperature microoscillation research, it is more convenient to employ higher-lag instruments for the determination of the latter.

Chapter 6

VARIATION IN THE VELOCITY OF THE WIND IN THE
LOWEST ATMOSPHERIC LAYER

The wind, i.e., the movement of the air in a definite direction along the terrestrial surface, is abated as it approximates the latter, this being due to friction and the formation of vorticity. The abating of the wind is determined, on the one hand, by the irregularities of the surface of the soil and its cover (by its "roughness"), and, on the other hand, by turbulent exchange. Due to turbulent intermixing, individual particles of air having a definite velocity along the terrestrial surface, descend into the lower layers, imparting part of their velocity to the lower lying air, and, vice versa, the lower particles, which lost their directional velocity, penetrate upward, decreasing thereby the wind velocity in the upper layers.

The process is complicated by the fact that the energy of the turbulent intermixing itself is fed at the expense of the kinetic energy of the wind. This circumstance must have an effect on the configuration of the wind in the zone of vorticity formation, i.e., within the range of heights of the irregularities of the underlying surface. In an exposed level locality these heights are measured in centimeters (the height of the grass, the unevenness of the soil), while under conditions of interlaced terrestrial relief, in sheltered localities, the configuration of the wind is disturbed in a layer, measuring in meters and tens of meters.

The wind is characterized by gustiness. An increase in its

velocity -- gusts -- alternates with an abating down to calm. The direction of the wind also changes continuously. The gustiness of the wind is a result of the vortices of various dimensions which are formed at the active surface and determine the degree of the turbulent intermixing of the air. Therefore, one of the methods for the determination of the turbulence factor is based specifically on calculating the gustiness of the wind.

The gustiness of the wind is tied in with the degree of stability of the atmosphere. This stability is increased in the presence of superadiabatic temperature gradients and is reduced in the presence of a temperature inversion. The irregularities of the underlying surface, too, induce an increase in the gustiness of the wind.

Gustiness, known otherwise as microoscillations of the wind in time, lead to its variability in space to the extent that at separate moments at a distance of only several meters apart under perfectly similar conditions the variation in the velocity of the wind may amount to 5 - 10 meters/per second.

Due to the gustiness of the wind, individual observations read by low-lag instruments result in random, non-characteristic values. This is the specific reason for using the mean wind velocities for definite time intervals (2 minutes by the Wilde anemoscope, 10 minutes by hand anemometer) for the comparative evaluation of the wind cycle.

For characterizing the variation in wind velocity with height, we will analyze Figure 31.

log in cm

Warm half of year

10 m/sec

Figure 31. Variation in wind with height in air layer from 2.5 centimeters to 258 centimeters.

I - temperature inversion; II - adiabatic gradient; III - super-adiabatic gradient; 1 - as per observations by Best and Gel'man; 2 - as per the D. L. Laykhtman formula.

The above figure characterizes the configuration of the wind during the warm half of the year as applicable to the middle latitudes for exposed level localities in the presence of super-adiabatic gradients (III), in the presence of a state of temperature equilibrium (II), and in the presence of a temperature inversion (I). The configuration is plotted in semilogarithmic coordinates and embraces the air layer from 2.5 to 258 centimeters. During the state of equilibrium, the configuration of the wind is a straight line, which indicates a logarithmic increase in wind velocity with height in the presence of which the wind velocity gradient is in inverse ratio to height. All this confirms the fact that the very

lowest point in the wind profile, in this case, lies above the height of the irregularities of the soil, and that, throughout the entire configuration of the wind, turbulence increases in direct ratio with height.

In the presence of a temperature inversion (Figure 31, I), the configuration of the wind, plotted in semilogarithmic coordinates, shows a clearly evident curvature, which indicates a non-uniform decrease in wind gradients with height. Up to the height of 2 meters, velocity increases slowly with height, but beginning with the height of 2 meters, the wind velocity increases at a more rapid rate than would follow from the logarithmic distribution, and already at the height of 50 - 100 meters the nocturnal wind velocities exceed those prevailing during the day.

Such an increase indicates that, under these conditions, the decrease in wind velocity gradients with height proceeds at a slower rate than under conditions of a state of equilibrium, which is perfectly in accord with the lower turbulence (see printed page 58 for analogous considerations pertaining to the temperature configuration).

The slower rate of increase in wind velocity up to a height of 2 meters, in the presence of a temperature inversion, can be explained in the following manner.

The film of cold, heavier, and less mobile air, which is forming in the presence of radiation chilling and stable stratification, covers all irregularities, and, becoming heavier on account of the run-off, equalizes them. The smoothening of the irregularities,

or the "roughness", decreases the expenditure of the kinetic energy of the wind for friction and vortex formation, by the virtue of which the decrease in the wind velocity in nearing the surface of the soil (or, which is the same, increase in wind velocity with height) is relatively abated. Above the 2-meter height a slower rate of increase in turbulence asserts itself, which causes the already indicated reverse effect, i.e., the relatively more rapid increase in wind velocity with height.

In the presence of super-adiabatic temperature gradients (Figure 31, III), the variation of the wind velocity with height is close to a logarithmic yet manifests some curvature when plotted in semilogarithmic coordinates. Below, up to the same height of 1.5 - 2 meters, the increase in velocity proceeds somewhat more intensively than it does in the presence of a state of equilibrium, and, beginning with that height, at a somewhat abated rate. The abatement in the increase of wind velocity, beginning with the height of 2 meters, is due to the more intensive growth of turbulence. The more abrupt abatement in the velocity of the wind from the indicated height as it nears the surface of the soil is due to the greater expenditure of the kinetic energy of the wind for friction and vorticity formation, in the presence of super-adiabatic gradients, as compared to a state of equilibrium and, particularly, to a state of temperature inversion.

The indicated considerations on the increase of the amount of kinetic energy of the wind expended on turbulence and friction, in the transition from a state of temperature inversion to superadiabatic gradients, receive their confirmation in the decrease of the mean

daylight velocity of the wind in the air layer up to 1500 meters, as compared to the nocturnal wind velocities.

Let us remind our readers that the increase in wind velocity from night to day, which occurs up to the 50 - 100-meter air layer, is due to an increase in the turbulent exchange with the above lying layers, where greater velocities prevail. The daylight decrease in wind velocities above the 50 - 100-meter air layer is determined not only by the intensive intermixing with the lower layer, where abated wind velocities prevail, but also by a general diminution in the kinetic energy of the wind, which is due to an increase in friction and in turbulence.

In characterizing the vertical variations of the wind, it is more convenient to use not the wind velocity gradients at various heights, but their ratios, since the gradients vary in relation to wind velocity, while the ratio of velocities, in the first approximation remains constant for a given height interval.

The computations for the variation of wind velocity with height during overcast and windy weather, i.e., when there is a state of equilibrium, can be made by the logarithmic formula as follows

$$u_n = u_1 \frac{\lg z_n - \lg z_0}{\lg z_1 - \lg z_0} \quad (15)$$

where u_n is the velocity of the wind at the height z_n ; u_1 is the known velocity at height z_1 ; z_0 is the index of roughness of the underlying surface.

This formula is applicable only in the case of exposed level localities. z_0 represents the height at which the velocity of the wind, i.e., the directional movement of the air, equals zero, assuming that the logarithmic regularity is preserved up to the very height z_0 . Indeed, as was indicated before, the configuration of the wind becomes deformed in the layer of roughness. In addition, the height of zero velocity of the wind, i.e., a directional current, unconditionally varies in relation to the thermal stratification of the lowest air layers. Nevertheless, the value of z_0 , at the below indicated method of its determination, is well harmonized with the nature of the underlying surface and is stably sustained with stability in the latter.

When the velocity of the wind at two heights in overcast windy weather is known, z_0 can be computed with the aid of the following formula derived from formula (15):

$$\lg z_0 = \frac{\lg z_n - \frac{u_n}{u_1} \lg z_1}{1 - \frac{u_n}{u_1}} \quad (16)$$

z_0 can also be determined graphically by plotting the configuration of the wind in semilogarithmic coordinates, i.e., by laying off the wind velocities along the abscissa in the conventional linear scale, and the heights in millimeters -- along the ordinate in logarithmic scale, as it is done in Figure 32.

log in mm

Meters/second

Figure 32. Determination of z_0 by the use of the graphic method.

Continuing the plotted straight line until it crosses the ordinate, we determine the height z_0 . For the wind configuration A, we have $\log z_0 = 1.7$, consequently $z_0 = 50$ millimeters = 5 centimeters. For wind profile B we have, respectively, $\log z_0 = 0.3$ and $z_0 = 2$ millimeters.

Numerous computations for the value of z_0 have shown that in exposed level localities with a grass cover the value of z_0 fluctuates, in relation to the height of the grass cover, within the range of 1 to 5 centimeters. Over a snow cover, the value of z_0 is reduced sharply to 0.05 and even 0.01 centimeter. Over a denuded and smooth soil, the value of z_0 is one centimeter.

It can be seen from Figure 32 that the greater the value of z_0 , the sharper the increase in wind velocity with height. But, with the same velocity aloft, the velocities at the lower heights will increase with the decrease in the value of z_0 .

Table 27 depicts the wind velocities at the heights of 2, 0.5, and 0.1 millimeters (at terraces with different values of z_0), in the presence of a wind velocity of 20 meters per second at the height of 10 meters, derived at by computation from the logarithmic formula and confirmed by actual observation.

TABLE 27

Velocity of wind (in meters/per second) at various heights, in the presence of a velocity of 20 meters/per second at the height of 10 meters

Nature of the soil cover	Z_0 in centimeters	Height (in meters)		
		2	0.5	0.1
Grass	3	14.4	9.3	4.0
Denuded soil .	1	15.4	11.4	7.0
Snow	0.05	16.0	14.0	10.6

Over snow-covered soil the velocity of the wind, in passing from the height of 10 meters to the height of 0.1 meter, is reduced to one half only, while over grass-covered soil it is reduced 5 times. It should be pointed out that a value of $Z_0 = 3$ centimeters is usually observed over a grass cover 10 centimeters tall.

As already indicated, the vertical variation in wind velocity is tied in with thermal stratification. Figure 33 depicts the ratios of wind velocities at 5- and 1-meter heights ($R_{5,1}$), in relation to the vertical temperature gradient [the temperature differential at the heights of 20 centimeters and 150 centimeters ($\Delta t_{20,150}$)] and to the wind velocity at the height of 2 meters (u_2). The graph is constructed by the day and night round-the-clock observations covering a period of 2.5 summer months at an exposed level area in the steppe. The grass covering the observation terrace was 10 centimeters tall, the value of $Z_0 = 3.4$ centimeters. By drawing in the isopleths, we smoothen the individual departures,

considering the latter a result of an error in the mean values due to a small number of observations (the initial data are given directly in figures).

Inversion Super-adiabatic gradient

Figure 33. Variation in the ratio of wind velocity at the heights of 5 meters and 1 meter ($R_{5,1}$), in relation to wind velocity u_2 and to the vertical temperature gradient Δt (0.2; 1.5). The lower figures designate the number of observations.

With a mild wind ($u_2 = 2$ meters/ second), the ratio varies in relation to the value of Δt within the range of 1.95 to 1.35. With the increase in wind velocity, the differences $R_{5,1}$ are reduced, and with a wind velocity of over 5 meters/per second, the difference $R_{5,1}$ does not exceed several hundredths. In the presence of a superadiabatic gradient, the value of Δt does not tell too sharply, but its effect is preserved in the presence of high wind velocities. When $\Delta t \approx 0.3$, $R_{5,1}$ does not depend on the velocity of the wind, retaining the same value -- 1.48.

If, during the day, in the presence of superadiabatic gradients, the departure of the vertical configuration of wind

velocities from the logarithmic is relatively insignificant, it is rather considerable at night, in the presence of a temperature inversion. Therefore, the logarithmic formula (15) may be used for the computation of the daylight vertical variation in the wind velocity in first approximation, but is certainly not applicable for the computation of the nocturnal variation.

D. L. Laykhtman suggested the following exponential formula for the configuration of the wind:

$$u_z = u_1 \frac{z_n^a - z_0^a}{z_1^a - z_0^a} \quad (17)$$

where u_z is the unknown wind velocity at the height z ; u_1 is the known wind velocity at the height z_1 ; z_0 is the index of roughness; a is the exponent determinable by the stability of the atmosphere.

When $a = 0$, formula (17) is automatically converted into formula (15).

The D. L. Laykhtman formula is most convenient for the interpolation and extrapolation of the results of direct observations of wind velocities at any two points to the entire near-the-surface air layer (up to 20 - 30 meters). The value of z_0 is determined by the above indicated methods. The wind velocities at two levels and the parameter z_0 make it possible to compute. Knowing a , the wind velocities at any level can be computed. To facilitate computations, the curve, linking the ratio of

velocities at two levels (R) and α , at a given value of z_0 , is preliminarily computed and plotted. When any value of R is obtained by observation, the value of α is determined by the curve, after which the wind velocities at any level are computed. The same curves can be utilized for the solution of the reverse problem, i.e., for the determination of the velocity ratios, by and, consequently, the determination of the velocity itself, if the other velocity is known.

Let us cite an example. Observations made at the heights of 5 and 2 meters show, respectively, wind velocities of 2.8 and 2.0 meters/per second, with $z_0 = 3$ centimeters. The ratio of the velocities $R = \frac{u_5}{u_2} = 1.40$. In Figure 34, constructed for a value of $z_0 = 3$ centimeters, we find that when $R_{5;2} = 1.40$, $\alpha = 0.25$. By the same graph we determine that when $\alpha = 0.25$, $R_{20;2} = \frac{u_{20}}{u_2} = 2.24$. Since $u_2 = 2.0$ meters/per second, $u_{20} = 4.5$ meters/per second.

The velocity of the wind at various heights can be computed by the same graph, in the presence of a logarithmic distribution, i.e., in the presence of a state of equilibrium, when $\alpha = 0$.

Figure 34. Wind velocity ratio R at indicated heights and the wind velocity at the height of 2 meters, with different values of α ($z_0 = 0.03$ meters).

In the presence of a logarithmic distribution of wind velocity, ^{the} correlationship of velocities at the heights of 20 and 2

meters equals 1.57. Consequently, if the velocity of the wind at the height of 2 meters equals 2 meters/per second, it will, at the height of 20 meters, be equal to 3.1 meters/per second. By comparing this numerical value with the one derived before, we obtain visual proof of the extent, to which the distribution of the wind can differ from the logarithmic and, consequently, the extent to which the exponential formula by D. L. Laykhtman introduces precision into the computation.

Since the determination of the vertical wind velocity ratio for the calculation of α is at times difficult, the empirically established ratio between α , on the one hand, and thermal stratification $\Delta t_{20,150}$ and wind velocity, on the other hand, may be utilized.

Table 28 cites the values of α , in the presence of various Δt and wind velocities at the 2-meter height.

TABLE 28

α in relation to u_2 and $\Delta t_{20,150}$

u_2 \ Δt									In the presence of a snow cover
	-1.5	-1.0	-0.5	0.0	0.5	1.0	1.5	2.0	
1	0.35	0.30	0.20	0.10	0.00	-0.05	-0.10	-0.10	0.20
2	0.25	0.20	0.15	0.05	0.00	-0.05	-0.05	-0.10	0.10
3	0.15	0.15	0.10	0.05	0.00	-0.05	-0.05	-0.05	0.05
4		0.05	0.05	0.00	0.00	0.00	-0.05	-0.05	0.05
5			0.00	0.00	0.00	0.00	0.00	0.00	0.00

The above data was tabulated from the result of continuous observation over a period of about one year in several localities of the USSR, yet the characterization of the values of α in relation to Δt is only an approximation. The reason why, on the average, the zero values of α are timed not to the zero value of Δt , i.e., to the state of equilibrium, but to $\Delta t = 0.5$, i.e., mild superadiabatic gradients, remains obscure, and requires additional research.

Substantial deflections of the exponent of α from zero are only observed in the presence of wind velocities, not in excess of 2 meters/per second, and predominantly during a temperature inversion. In the presence, at the 2-meter height, of wind velocities from 4 meters/per second and over, the values of α oscillate within the range from +0.05 to -0.05. It is not difficult to see from Figure 34 above that the ratio of wind velocities does not substantially differ from the logarithmic, which, of course, facilitates computation considerably. Actual observation shows that, in windy and overcast weather, the logarithmic distribution of wind velocity with height is preserved up to 100-meter level.

In the last column of Table 28 are cited the values for α in the presence of a snow cover. In this case, as confirmed by numerous investigations, the values of α vary not in relation to t , but in relation to the velocity of the wind.

The change of the algebraic sign of α , as between day and night, indicates that the vertical variation in the mean diurnal

wind velocities should be nearer the logarithmic than the vertical variation for the daylight and nocturnal hours taken separately. This forms the basis for the application of the more simple logarithmic formula to the approximated calculations of the mean diurnal wind velocities at various heights.

The approximation formulas below are intended for the calculation of the mean diurnal wind velocities at various heights within the layer from 2 to 30 meters by the reading of the anemoscope located at a height of 10 meters over exposed level ground. They are derived from formula (15) by substituting the values of the logarithms z and z_0 . For the warm part of the year, the value of z_0 for the entire profile is accepted as equal to 3 centimeters, and for the time of the year, when the snow cover is on the ground, when calculating down from the height of 10 meters, it is accepted that $z_0 = 0.05$ centimeters.

To calculate the wind velocity up from the 10-meter height the formula immediately below can be used at all times:

$$u_z = u_{10} (b + 0.6), \quad (18)$$

where u_z is the unknown wind velocity at height z ; u_{10} is the wind velocity, by the anemoscope reading, at the height of 10 meters; b is the variable term that changes in relation to height z for which the wind velocity is being determined.

For calculating down from the 10-meter height, this formula can be used only in the absence of the snow cover. In the presence of a stable snow cover, which smoothenes all irregularities, it is necessary to use another formula when calculating down from the

10-meter height, namely:

$$u_z = u_{10} (c + 0.8), \quad (19)$$

(designations are the same as in the preceding formula).

Table 29 below furnishes the values of the variables b and c.

TABLE 29

The values of b and c for the calculation of
wind velocities at various heights

	Height in meters					
	2	3	5 - 7	8 - 14	15 - 24	25 - 30
b	0.1	0.2	0.3	0.4	0.5	0.6
c	0.05	0.1	-	-	-	-

The calculations of wind velocities at various heights by the simplified formulas (18) and (19) and also by the precise formula, but with indirect determination of u_{10} by Table 28, prove their usefulness in the determination of mean wind velocities, let us say, for a 10-day period or a month. To calculate wind velocities at individual moments of time, particularly at night, it is best to use the exponential formula (17), with the determination of u_{10} from direct observations.

On the basis of the above indicated regularities in the variation of wind velocities with height in the lowest atmospheric

layer, it becomes possible, in an exposed level locality, by the anemoscope readings, to calculate the annual variation in wind velocity at various heights, for day and night, at various latitudes, taking into consideration zonality (see printed page 55), variations in Δt and also z_0 to the extent to which it is determined by the snow cover.

Bearing in mind the well known minute geographic variability of wind velocity at the anemometer level (in the territory under our observation), we will now analyze some observation data from two extreme zones: the northern forest zone ($\phi = 65^\circ$) and the southern steppe zone ($\phi = 45^\circ$). By so doing we temporarily exclude from the analysis some characteristics of the intermediate latitudes determinable not as much by general climatic conditions as by the peculiarities of the landscape covered with sylvan vegetation. For convenience the data is presented in Figure 35 by isopleths. The isolines for the lowest half-meter layer are shown by dotted lines since at the very surface of the earth the roughness of the soil cover z_0 exerts a considerable effect. In calculations for latitude 45 degrees north, the value of z_0 for the entire year is accepted as equal to 3 centimeters. For latitude 65 degrees north for the winter months when there is a stable snow cover, the value of z_0 is 0.05 centimeters, while for the rest of the year it is 3 centimeters. The entire characteristic below pertains to an exposed open place.

Day Night

Height Height

Figure 35. The annual variation of the configuration of the wind (in meters/per second) in the lowest 10-meter air layer.

During the daylight hours there are no substantial differences between the north and the south with the exception of the winter months when in the north, above the snow cover, a 3 meters/per second wind is observed in the lowest half-meter air layer and even at the 10-centimeter level the velocity of the wind is in excess of 2 meters/per second. With the disappearance of the snow cover this diversity disappears and in the north as well as in the south the wind velocity in the air layer from 0.1 to 10 meters varies from 1 to 5.5 meters/per second with the wind velocity from 1 to 4 meters/per second prevailing in the lowest 2-meter air layer. At the 1-meter level, the mean daylight wind velocity is about 3 meters/per second.

We will now analyze the mean values of the nocturnal wind velocities. The nocturnal velocities manifest a clearly pronounced annual variation and, in addition, they differ substantially in geographical cross section. During the transition seasons these

diversities are somewhat smoothened. During the winter they embrace the air layer directly at the edge of the earth's surface, and they are determined by the presence, in the north, of a stable snow cover which is absent in the south. During the summer they extend throughout the entire contemplated air layer and are tied in with a more sharply pronounced temperature inversion in the south.

Table 30 cites the nocturnal wind velocities at the 0.5 meter level.

TABLE 30

Nocturnal wind velocities at the 0.5 meter level

Latitude	I	III	V	VII	IX	XI
65°	3.1	2.7	1.5	1.3	1.8	2.5
45°	2.1	1.8	1.2	0.8	1.3	2.1

At the 10-centimeter level the mean nocturnal wind velocity during the summer months drops to several tenths of a meter per second.

The comparison of the daylight and the nocturnal wind velocities make it possible to judge the diurnal range of wind velocity and its annual as well as latitudinal variation. At the 0.5 meter level the difference between daylight and nocturnal wind velocities during the summer is 1.0 meter per second in the north, and more than 1.5 meters per second in the south. During the winter, in the north, there is no diurnal wind velocity range, while in the

south, the range is in excess of 0.1 - 0.3 meters/per second.

It is important to emphasize that, in the lowest atmospheric layer, the diurnal range of wind velocity, during the warm part of the year, is greater than its annual range as can be seen from Table 31 where the correlationship of the daylight wind velocities u_d and the nocturnal wind velocities u_n is taken as the index of the diurnal range, and the correlationship of mean wind velocities for January u_I and for July u_{VII} is taken as the index of the annual range.

TABLE 31

Characteristics of the diurnal $\left(\frac{u_d}{u_n}\right)$ and the annual $\left(\frac{u_I}{u_{VII}}\right)$

range of wind velocity at various heights

Latitude	Height in meters	Diurnal range u_d / u_n						$\frac{u_I}{u_{VII}}$
		I	III	V	VII	IX	XI	
65°	10	1.00	1.19	1.44	1.42	1.31	1.02	1.17
	2	1.00	1.18	1.56	1.59	1.86	1.03	1.33
	0.5	1.00	1.18	1.60	1.77	1.44	1.04	1.72
55°	10	1.05	1.25	1.92	2.16	1.73	1.08	1.87
	2	1.03	1.28	2.20	2.50	1.88	1.11	1.55
	0.5	1.04	1.40	2.34	2.72	2.10	1.12	2.04
45°	10	1.06	1.25	1.45	1.86	1.37	1.04	1.31
	2	1.09	1.34	1.62	2.38	1.48	1.09	1.32
	0.5	1.14	1.39	1.83	3.11	1.84	1.10	1.36

By studying the correlationship of the diurnal and annual ranges of wind velocities, we arrive at the conclusion that, if the annual variation of the wind is taken into account in the characterization of the wind cycle, the diurnal variation of the wind during the summer months is of still greater importance.

When analyzing individual air layers for prevailing wind velocities, it becomes necessary to segregate the lowest 2-meter layer as a layer where a wind velocity of less than 3.5 meters/per second prevails. During the summer, the mean nocturnal wind velocities are less than 2 miles/per second. In the 2 - 10 meter layer, during the day, the prevailing velocities are 4 - 5 meters/per second, and, during the summer nights, from 2 - 3.5 meters/per second in the north, and from 1.5 - 3 meters/per second in the south.

The above cited values are, of course, to be considered as approximations only, yet they lead to one more conclusion of practical importance, namely: the vertical zonality of wind velocity in the lowest 10-meter air layer is an existing reality which must be taken into account both in theoretical calculations as well as in the planning of the activities of agricultural production, communal building programs, climatic therapy, etc. It must also be remembered that, as was indicated above, the ratios of the wind velocities at various heights during the day are comparatively stable and become even more so with an increase in velocity. Therefore, although the increase in the mean wind velocities, in passing from the 1 - 2 meter layer to the 5 - 10 meter layer, from 3 to 4.5 meters/per second (1.5 times) may not seem substantial, yet it means that in a storm, when the wind

velocity at the 5 - 10 meter layer is 20 meters/per second, it is only 13 miles/per second at the 1 - 2 meter layer. This difference is, no doubt, substantial.

The peculiarities of a locality exert a considerable effect upon wind velocity (a special chapter is devoted to this problem).

SECTION III

THE MICROCLIMATE OF THE SOIL AND THE ADJACENT AIR LAYERS

Chapter 7

The Temperature of Denuded Soil (Black Fallow) During the Warm Season

In the upper soil layer the vertical temperature gradients are even greater in value than those in the air. The horizontal soil temperature differentials, in relation to its characteristics and cover, are also quite considerable. Therefore, the climate of the upper layers of the soil must be considered as a microclimate.

The underlying causes for the high temperature gradients in the top layer of the soil are different from those prevailing in the air. In the air, as was indicated above, the decisive part is played by the rising values of turbulent exchange with height, while in the soil, due to its greater volumetric thermal capacity (more than 1000 times greater than that of the air), the temperature gradient is basically reduced with depth because the greater part of the heat is absorbed by the upper layers, and, consequently, the thermal current is substantially weakened with depth. Only in dry weather a certain influence on the increase of the gradients in the very top layer of the soil is exerted by the decrease in its temperature conduction due to lower density and lower moisture content.

The temperature variation of the soil with depth, under

desert conditions, as observed on a summer day, is represented in Figure 36.

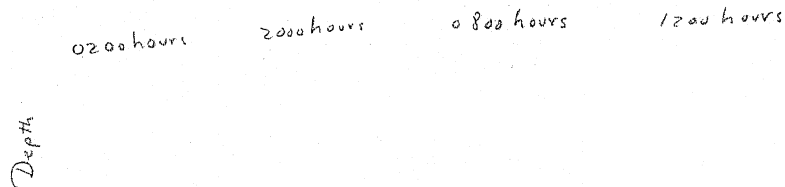


Figure 36. Tautochronograms of the soil temperature as observed at Arys' on 29 July 1945.

The temperature jump in the top 5-centimeter layer at 1200 hours is over 20 degrees Centigrade while, at the same time, in the 5 to 10-centimeter layer, the temperature differential is only 4 degrees Centigrade, and in the 15 to 20-centimeter layer it is only 0.2 of a degree.

At 1200 hours the temperature maximum prevails at the surface of the soil and is diminishing with depth. At 0200 hours (at night), on the contrary, it is the temperature minimum that prevails on the surface of the soil and, with depth, the temperature is increased. The temperature variation becomes more complex during the transitional hours. At 0800 hours the temperature minimum is observed in the intermediate layer A, with the temperature rising in an up and down direction from the latter, and at 2000 hours, the temperature maximum is observed in the intermediate layer B, and it is dropping in an up and down direction from it.

The qualitative characteristics of the vertical temperature distribution, as represented in Figure 37, are preserved through the warm season of the year under any climatic conditions, with only

the transitional hours shifting in relation to the rising and setting of the sun and showing a considerable variation in the temperature level.

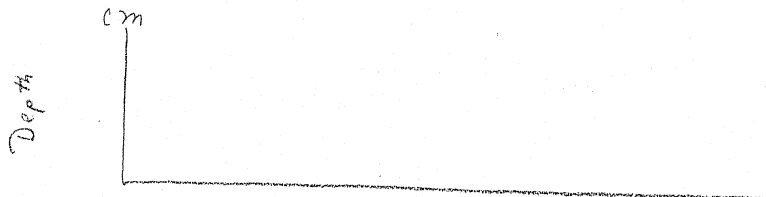


Figure 37. Tautochronograms of soil temperature (black fallow) in the temperature deflections at the 10-centimeter depth, as observed at Ovtino, May 1937, at 1300 hours.

1 - monthly mean; 2 - 15 May, overcast, after rain; 3 - 20 May, stable clear weather.

However, on individual overcast rainy days, isothermy may be observed in the soil even during the near-midday hours. This can be seen in Figure 37 above (observations taken in the environs of Leningrad), in which the vertical mean temperature variation in the temperature deflections, at the 10-centimeter depth, for the month of May 1937, is represented, for the cases of stable clear weather and overcast after a rain.

During the winter, the rising of temperature with depth predominates.

The diurnal temperature range, as well as the annual range, gradually disappear with depth. The specific depth at which it occurs is determined both by the temperature conduction^(c) of the soil as well as by general weather conditions and the nature of the soil cover, which factors cumulatively determine the value of the thermal

current from the surface into the depth of the soil, or the value of the reversed current.

The microclimate of the soil of a field, a meadow, and a black fallow parcel of land have great practical significance.

From the moment of sowing to the moment of germination, and even for some time after, while the grass cover is insignificant, the temperature of the soil of a sown field is essentially the same as the temperature of black fallow, i.e., denuded turned-up soil.

Soil temperature observations at meteorological stations were conducted in the past almost exclusively under the so-called "natural surface".

The very term "natural surface" is extremely indefinite: the natural vegetation of a meadow, the specially sown and regularly mowed down meadow grasses, weeds of any degree of denseness, and, finally, the soil simply denuded (by trampling or drought) of all vegetation, all these may be relegated, with the same degree of logic, to the classification "natural surface". The data cited in the existing climatological manuals specifically refer to this "natural surface", which circumstance makes their utilization difficult.

During the recent years, soil temperature observations are made under a denuded surface. The results of these observations can be utilized for the analysis of the conditions of seed germination.

The widely shared opinion among practical agricultural men that, during the spring, argillaceous soils are cold while arenaceous soils are warm is corroborated by these observations. As per the data collected by E. P. Arkhipova, the difference between the temperatures of the top layers of arenaceous and argillaceous soils, under the spring and summer conditions of the northwest forest zone of the European territory of the USSR, is about 1.0 - 1.5 degrees Centigrade in favor of the arenaceous soils (see Table 32). Only in the autumn, in September, the argillaceous soil becomes warmer than the arenaceous.

TABLE 32,

Mean monthly temperature of arenaceous, subarenaceous and argillaceous soils under black fallow

	V		VI		VII		VIII		IX	
Soil	depth in centimeters									
	5	20	5	20	5	20	5	20	5	20
Arenaceous	11.0	9.5	16.0	15.0	20.0	18.0	18.0	18.0	9.5	10.5
Subarenaceous	10.5	9.0	15.5	14.0	19.0	17.5	17.5	17.0	10.0	11.0
Argillaceous	10.0	8.5	15.0	13.5	18.5	17.0	17.0	16.5	10.5	11.0
Air at height of										
200 centimeters	9.0		14.0		17.0		16.0		9.0	

Such a ratio of temperatures is determined by the greater thermal capacity of argillaceous soils and also by the fact that argillaceous soil, being nonpermeable to water, promotes the vaporiza-

tion of the latter as a result of which the latent heat of vaporization from argillaceous soils is greater than same from arenaceous soils, leaving less heat for the warming of the soil.

The greater thermal capacity of subargillaceous and argillaceous soils is determined by their greater moisture capacity and greater water content. The latter circumstance is illustrated in Table 33 in which the dampness of soils of various mechanical composition under conditions of the forest zone is cited (by the data collected by A. K. Filipova).

TABLE 33

Dampness and field moisture capacity of soils in percentage
of absolute dry weight of soil in the
0 - 20-centimeter layer

	Months	Soil		
		arenaceous	subargillaceous	argillaceous
Dampness	IV	20	30	45
	V	10	20	40
Field moisture capacity		35	40	50

Special research (Table 34) on the effect of the mechanical composition on the thermal cycle of the top layers of the soil were conducted by Kreutz in Germany (University of Hessen)

TABLE 34

Mean monthly temperature of the soil at the depth of
5 centimeters (according to Kreutz)

Months	Arenaceous			Argillaceous			Humus		
	Maximum	Minimum	Range	Maximum	Minimum	Range	Maximum	Minimum	Range
V	19.4	8.2	11.2	17.5	8.8	8.7	17.8	10.2	7.6
VI	19.8	12.3	7.5	18.5	13.0	5.6	18.9	14.4	4.5

Comparing the thermal cycle of the arenaceous (fraction 0.5 millimeter - 70 percent), argillaceous (same fraction 4 percent), and humus soils (same fraction 34 percent, humus 24 percent) under a denuded surface we can see that the greatest warming during the maximum period and the greatest range are, indeed, observed in arenaceous soil.

In argillaceous soil, the temperature range decreases mainly at the expense of the reduction in the maximum value which circumstance serves as an indirect index of the predominant role of vaporization which reduces the daylight temperatures. The diminution of the temperature range in humus soils is determined not only by a reduction in the maximum value but also by a rise in the minimum value which indicates, first, the leading role of lower temperature conduction to be manifested to an equal degree by day and by night and, secondly, the possible rise in daylight temperatures due to the lower albedo of the dark-red humus soil.

Of the three types of soil studied comparatively by Kreutz, the arenaceous soil is the warmest and the argillaceous soil is the coldest in their mean May values. The argillaceous soil remains the coldest also in June.

Thus, the availability in the spring of warm and cold soils can be considered as physically substantiated and experimentally corroborated. The easily drainable and relatively dry soils are the warm soils; the heavy superdampened soils are the cold soils. It must be emphasized that this regularity is observed only during the spring and the summer. During the cooler period of the autumn when the light and heavy soils are equally superdampened, the thermal cycle of the soils is equalized and arenaceous soil may be found to be even colder than argillaceous, particularly at night. Due to its high thermal conduction it chills more rapidly.

Still greater differences are observed as between mineralized soils and peat soils.

TABLE 35

Soil temperature in mineralized and peat parcels of
Khibina (Kola Peninsula), 21 - 31 May 1934

Soil	Depths in centimeters			
	5	10	15	20
Mineralized	15.3	13.2	12.1	10.8
Bog-peat	11.1	10.1	9.0	8.3
Differential	4.2	3.1	3.1	2.5

As seen from Table 35, compiled by the Agro-hydro-meteorological Institute from the works of N. A. Kurmangalin, peat soil at the beginning of the vegetation period is substantially colder than mineralized soil. The slow warming of peat soils is also determined not only by their low heat conduction, but by the relatively high value of vaporization rising from them.

Cultivating the soil which, to a considerable extent, is reduced to turning it up or consolidating it, has an effect on the rate of warming. Experience shows that there exists a definite optimum of porosity, a deflection from which in one or the other direction, leads to a slower rate of warming of the soil. Apsits gives (in Table 36) the characteristics of the spring thermal cycle of the soil in relation to its porosity (as observed at the Experimental Station of the Latvian University).

TABLE 36

Soil temperature at the depth of 15 centimeters
on 18 - 26 May 1933

	Parcels			
	I	II	III	IV
General porosity	33	46	49	63
Temperature	11.2	11.7	11.5	11.3

The more rapid spring warming of rolled soils in the presence of considerable porosity is cited by Gerasimov who bases himself on the observations of the Perm' Agricultural Institute (Table 37).

TABLE 37

Soil temperature at the depth of 10 and 20 centimeters
for the period 30 April - 20 May 1937

Type of cultivation	Total porosity	Temperature at	
	(1 to 10 cm)	10 cm	20 cm
Soil cultivated in the spring before sowing:			
(1) With plow and harrow (single furrow)	61	4.5	3.9
(2) Additionally, with scraper	58	5.3	4.3

The optimum porosity, in relation to the rate of warming of the soil, is tied in directly with the maximum temperature conduction and is bound to vary in relation to the mechanical composition and the dampness of the soil.

The spring warming of the soil is promoted by the soil ridge-type cultivation of plants. According to Kurmangalin, who used the data of the Agro-hydro-meteorological Institute, the temperature of the soil ridges, in its mean value, is higher by one degree Centigrade than the temperature of level spots (Table 38).

TABLE 38

Soil Temperature in relation to the form of surface --
Khibina, 11 - 30 June 1934

Form of Surface	Depths in centimeters			
	5	10	15	20
Ridges	13.6	11.3	10.5	9.8
Level field	12.6	10.5	9.4	8.3
Differ- ential	1.0	0.8	1.1	1.5

A still greater effect on soil temperature is exerted by the very fact of disrupting the natural soil cover. The denuding of the soil of its natural grass cover and sod causes a sharp increase in its temperature, particularly during the day, as confirmed by the observations of Sverdlovsk observatory, depicted in Figure 38.

Figure 38. Soil temperature isopleths at the depth of 20 centimeters, Sverdlovsk, period 1901-1910.

A- denuded soil surface

B-soil surface under its natural cover.

Observations conducted by the sovkhos "Industria", located on the Kola Peninsula, have shown that the removal of sod during the summer raised the mean diurnal soil temperature by 1-3 degrees Centigrade (Table 39), with the sod having played the part of a heat insulator.

TABLE 39

Mean Diurnal soil temperature at the 5 centimeter depth

Soil	VII			VIII			IX
	1	2	3	1	2	3	1
Mineralized (No sod)	17.1	18.8	16.3	14.4	14.0	12.4	8.2
Mineralized (with sod)	13.9	17.6	15.6	12.8	12.9	11.8	8.1
Differential	3.2	1.2	0.7	1.6	1.1	0.6	0.1

The effect of sod manifests itself in a considerable part of the general climatic phenomena relating to the temperature of a soil with a natural cover, which phenomena induce a lowered temperature, by comparison with cultivated fields, particularly during the period of the spring temperature rise.

From Table 32 it follows that during the month of May in the forest zone of the northwest, the temperature of the upper layers of the soil under black fallow is, in its mean value, somewhat higher than the temperature of the air (the only exception being argillaceous soil at the depth of 20 centimeters). The proximity of the mean diurnal temperature of the top layers of the soil to the temperature of the air (within 1 - 2 degrees) is preserved throughout the entire European territory of the USSR, but the relative homogeneity of thermal conditions is only a homogeneity in the sense of its mean value. In reality, it is observed only during the morning and evening hours. For the rest of the diurnal period, the thermal cycle of the soil is different from that of the air, with the difference increasing in the direction toward the surface of the soil.

Figure 39 depicts the diurnal temperature variation in the soil at various depths during the month of May (in accordance with observations at Pavlovsk). The temperature contrasts, experienced by germinating plants, both, with relation to the diurnal cycle (the diurnal range at the surface is about 20 degrees Centigrade, and at the depth of 20 centimeters 3.5 degrees Centigrade) and with relation to depth.

During the first half of the day the parts of the plants at the surface of the soil have an environmental temperature that is by 8 - 12 degrees higher as compared with same for the parts of the plants that are located at the depth of 10-20 centimeters.

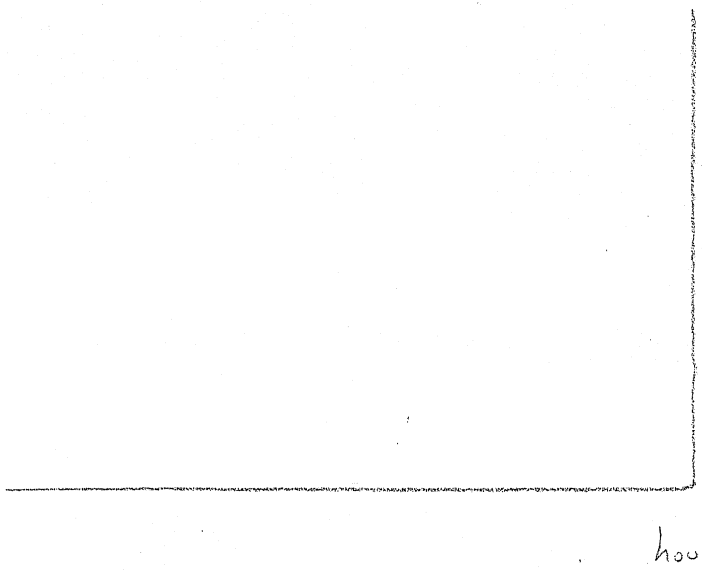


Figure 39. Diurnal temperature variation at various depths in arenaceous soil, Pavlovsk, during the month of May.

The conditions depicted in Figure 39 are the conditions of the forest zone. In more southerly areas, with prevailing clear weather and, consequently, a greater daylight thermal current into the depth of the soil, the temperature contrasts, naturally, increase, and on individual days, the vertical temperature variation in an arable stretch of land attains a value of over 20 degrees Centigrade.

In connection with this, the importance of the depth of

seeding, with relation to the germination of the seeds, is very obvious.

Of great practical importance is the temperature of the soil surface, particularly, since it differs substantially from the temperature of the air, with the exception of those cases, when the soil is shaded by grass or ligneous vegetation.

The mean maximum and minimum differentials between the maximum temperatures of the surface of the soil (black fallow) and the air at the 2-meter level, as observed by the Agro-hydro-meteorological Institute at the experimental localities near Leningrad (Ovtsino) and in the Northern Caucasus (the Maykop Experimental Station of the All-Union Institute for Plant Cultivation), are depicted in Table 40. Here and further below, the temperatures of the soil surface are given as registered by mercury and alcohol thermometers located at the surface.

TABLE 40

Differentials Δ of maximum temperatures in the surface of black fallow and air at the 2 meter level

Years	Ovtsino						Shantuk					
	V			VI			IV			V		
	Δ Mean	Δ Max	Δ Min	Δ Mean	Δ Max	Δ Min	Δ Mean	Δ Max	Δ Min	Δ Mean	Δ Max	Δ Min
1937	9	16	1	8	12	4	-	-	-	-	-	-
1938	*	6	12	6	9	16	7	-	-	-	-	-
1939	9	16	0	12	24	0	15	24	4	18	26	5

As seen from above Table, the surface of black fallow has a considerably higher maximum temperature than the maximum temperature than the maximum temperature of the air, with the differential being greater in the south. The daylight temperature of the soil surface is the greater than the temperature of the air, the greater the radiation balance, and the lesser the amount of heat that flows into the depth of the soil and into the air, both directly, as well as in a hidden manner (the latent heat of vaporization). Therefore, the greatest differentials between the respective temperatures of the soil surface and the air are observed in clear calm weather, in the case of dark-colored dry loose soils.

The mean variation in the maximum temperature differentials soil surface - air, with relation to the meridional solar altitude, on clear days, by the month, as per observations in Tbilisi and Pavlovsk (Leningrad) the soils being arenaceous, is presented in Figure 40.

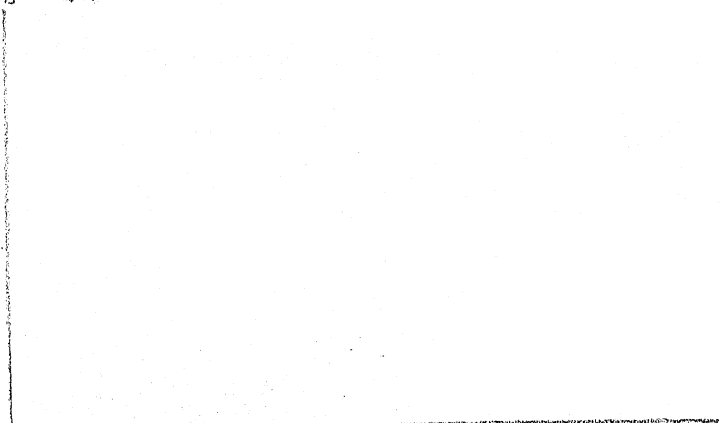


Figure 40. 1- Tbilisi (observed over the period 1880, 1888-1890).
2- Pavlovsk (observed over the period 1901-1910).

The solar altitude, in this case, is the index of the radiation cycle. Almost all points at both localities are lying in the respective straight lines, it being the case that the Pavlovsk line, with the same solar altitude, shows a lower temperature differential, which is fully explained by the greater soil dampness in Pavlovsk as compared to Tbilisi.

It follows from the analysis of Figure 40 that in clear summer weather the temperature of the soil surface is by 15- 25 degrees Centigrade higher than the temperature of the air. In individual cases the differential may be even greater, but the relation to solar altitude is still preserved. This makes it possible, when the maximum air temperatures are known, to make an approximated evaluation of the absolute temperature maximums of the surface of a demuded mineralized soil at various latitudes, by taking into account the annual variation in the meridional solar altitude.

The absolute temperature maximums, cited in table 41, are possible on the surface of a completely dry soil, demuded or covered with scarce non-shading vegetation, under conditions of sunshine. [Table 41 on next page]

The minimum temperature of the soil surface is of particular interest.

The minimum and, generally speaking, the nocturnal temperature differential between the soil surface and the air are lower in value and have a reverse algebraic sign (See Table 42), since, during the period of minimum temperatures the soil surface is colder than the air.

TABLE 41

Absolute maximum temperatures of soil surface, by months, for latitudinal zones (soil denuded during the summer and covered with snow during the winter).

Latitude	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
45°	30	40	50	60	65	70	70	70	60	50	40	30
50	20	25	40	55	65	70	70	65	55	40	30	20
55	0	15	30	50	60	65	65	60	50	40	20	15
60	0	0	25	45	55	60	60	55	45	35	15	10
65	0	0	0	40	55	60	60	55	35	30	10	6

The advection of cold air in overcast weather is an exception to this rule, in the case of which exception the temperature of the soil surface may be 1-3 degrees higher than the temperature of the air.

TABLE 42

Minimum temperature differentials between the soil surface and the air at the 2 meter level

Month	Pavlovsk (1896-1905) denuded soil			Months	Poltava (1910-1913), denuded turned-up soil		
	Δ Mean	Δ Extreme			Δ Mean	Δ Extreme	
V	-1	-5	3	IV	-3	-7	3
IV	-1	-5	3	V	-3	-9	1

In the mean, the minimum temperature differentials between the soil surface and the air increases somewhat toward the south, due to the increase in the recurrence of clear nights with intensive radiation chilling. However, the basic effect is exerted upon it by purely local conditions, first of all, by the degree of the soil consolidation. In loose freshly plowed soils, the minimum temperature is by 3-5 degrees lower as compared to adjacent consolidated soils. This specifically is the reason for the occasional formation of hoarfrost on black fallow, while it is absent from the adjacent well trampled patch.

Minimum temperatures on the surface of peat areas are in relation to the soil dampness. In the case of damp peat soils they may be higher than in mineralized soils, in the case of dry peat soils - lower.

Higher minimum temperatures (as compared with the air) of the soil surface can be observed systematically only in areas with local advection of cold air.

A good example of such local advection is the meteorological stage of the Tbilisi Observatory. As can be seen from Table 43, the minimum temperature differential between the soil surface and the air, in that case, has a negative sign only during the spring and autumn. During the summer months the differential is positive, both in overcast and in clear weather. This phenomenon can be explained as follows:

The Tbilisi meteorological stage is located in a garden. In the presence of radiation chilling, the surrounding tree top

surfaces are cooled down to a lower temperature on account of their lower thermal capacity. During the summer, when the daylight heat reserve in the soil is particularly great, the soil surface at night turns out to be not only warmer than the tree tops, but also warmer than the air, which is cooled by the foliage and subsequently flows off onto the stage.

TABLE 43

Mean differentials of the minimum temperatures between the soil surface and the air at the height of 2 meters on clear and overcast nights, Tbilisi, 1880, 1888-1890

Nights	III	IV	V	VI	VII	VIII	IX	X
Clear	-1.6	-1.1	-0.4	0.2	0.3	0.6	-0.4	-0.9
Over-								
cast	-1.0	-0.2	0.2	0.4	0.8	0.6	0.0	-0.6

A similar phenomenon can be observed at one of the black fallow parcels at the Maykop Experimental Station located at the bottom of a gulch, the slopes of which are overgrown with forest.

Chapter 8

The Microclimate of a Wheat Field During the Period of Tall Grass.

During the phase sowing-germination, as was already indicated, the microclimate of agricultural fields, does not essentially differ from the microclimate of black fallow, but with the progress of growth, the plants originally under the influence of climatic conditions, themselves begin to exert an ever increasing effect upon climatic conditions.

The microclimate of agricultural crops, or as it is sometimes designated, the phytoclimate, develops in direct relation to the plants proper, the total weight of the green mass per unit of area, the degree of density and shape of the foliage surface, the transpirational coefficient, i.e., in direct relation to everything that determines the penetration of solar rays into the depth of the vegetation cover, to turbulent exchange, and to the latent heat of vaporization.

Therefore, the microclimate of fields may vary extensively. It is to be regretted that not all the research in this field took the above into consideration, as a result of which a considerable portion of the microclimatic characteristics of various crops lack the clear indication as to what type of vegetation cover (height, formation, density, etc) is being referred to, which in turn, leads to contradictory conclusions.

While, in accordance with some data, the climate of a wheat field approximates that of black fallow and substantially differs from the meteorological conditions at the height of a

meters (in observation booth), others insist that the temperature in the midst of the grass cover of the very same wheat field differs but little from the thermometer readings registered at the height of 2 meters. In individual cases, these or the other may be right, yet without a clear indication as to the specific grass cover, to which the data has reference, the data cannot be summarized, and consequently, cannot be practically utilized.

Hence, only the type of microclimatic data, that is accompanied by the description of the grass cover, has scientific and practical value.

An integral characteristic of the grass cover, as a climate forming factor, is the weight of the vegetation mass. Parallel observations of microclimate and vegetation mass are available in the case of wheat, which makes it possible, by the example of this specific crop, to illustrate the diversity of the microclimate of herbaceous vegetation and its relation to the abundance of the grass cover, to the agro-technical methods employed, to the weather and general climatic conditions.

The maximum vertical air temperature variation in the air layer from the surface of the soil to the height of 200 centimeters, in the case of a wheat field, as observed in Yershov (area of Saratov) and Ovtstino (area of Leningrad), is depicted in Figure 41. For comparison purposes, almost in all observations, parallel data for black fallow and the characteristic of the grass cover in the form of the weight of the green mass (i.e., leaves, stems, spikes) or the weight of the harvest (to a

small harvest, all other conditions being equal, corresponds to less abundant grass cover).

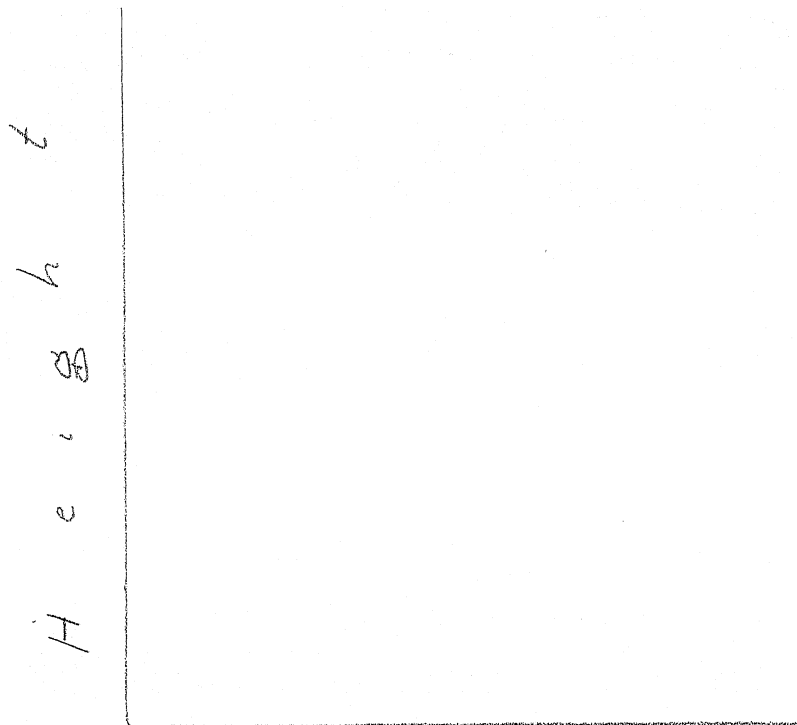


Figure 41. Maximum air temperature variation with height in departures from the maximum at the height of 2 meters.

- A- Ovtino, 1940; wheat; the spiking inflorescence phase;
- B- Wheat, lactescent ripeness phase - harvesting; I- wheat, greenmass, 3.0 kg/m²; II- wheat green mass, 1.2 kg/m²;
- III- black fallow; G - Yershov; wheat, inflorescence - lactescent ripeness; I-year 1935, harvest 10.8 centners per hectare; II-year 1933, harvest 4.0 centners/per hectare

By comparing the individual graphs, it becomes possible to establish that the less abundant the grass cover, the more the maximum temperature in its depth differs from the air temperature at the 2-meter height. This characteristic holds true for both Yershovo and Ovtino. In the presence of an abundant grass cover, isothermy is observed in the entire layer from the soil surface. In the presence of a grass cover of low abundance, the temperature is almost the same, as in the case of black fallow, and at times even higher.

As already indicated, the influence of the grass cover is determined by its effect upon the radiation balance, vaporization, and turbulent exchange.

In relation to the abundance of the grass cover, and also to solar altitude, the percentage of solar radiation, penetrating into the depth of the grass cover, may vary greatly. The greater the solar altitude, the greater, all other factors being equal, the percentage of direct solar radiation that will penetrate to the surface of the soil.

In the presence of an abundant grass cover, only a small part of the direct solar radiation penetrates to the soil, while most of it is absorbed by the grass cover at its various levels. This, specifically, is the condition that inhibits the formation of a clearly pronounced maximum at some definite level.

Still, the decisive effect is exerted by the combined latent heat of evaporation and transpiration, which are related directly to the abundance of the grass cover. The approximated characteris-

tic of the diurnal consumption of moisture by plants in the presence of various degrees of abundance of the grass cover (as observed by D. I. Shashko), is depicted in Table 44.

TABLE 44

Diurnal consumption of moisture by plants in the presence of various degrees of abundance of the grass cover.
(in millimeters)

Phase of growth	Evaluation of grass cover			
	Poor	Fair	Good	Abundant
Inflorescence				
lactescent				
ripeness	1.5	2.3	3.0	3.7

Converting into calories, we derive the diurnal consumption of heat as 80-90 calories/ per centimeter squared, in the case of a poor grass cover, and as 200-220 calories/ per centimeter squared, in the case of an abundant grass cover. Hence, in the first case, it will take a 100 more calories to warm the soil and the air than in the second case, which circumstance must show its effect upon the temperature gradients.

On the basis of the above, the conclusion can be made that, in the Stakhanov fields with abundant grass covers, the daylight air temperatures, both at the soil surface and at the

height of the spike, are practically the same as the air temperature at the height of 2 meters (at the observation booth). In a poor grass cover, the daylight temperatures are higher, particularly, at the very surface of the soil.

An example of the effect of the amount of vegetation mass upon the soil temperature is depicted in Table 45 (as per N. N. Yakovlev).

TABLE 45

Temperature of the soil in the presence of various amounts of vegetation mass. Summer wheat, tube formation phase.

Depths (in Centi- meters)	Weight of raw mass							
	380 grams				939 grams			
	Observation hours							
	0700	1300	2100	Δ 1300- 0700	0700	1300	2100	Δ 1300- 0700
0	19.5	31.9	19.0	12.4	18.9	28.1	18.3	9.2
5	18.1	27.1	21.6	9.0	17.8	24.3	20.7	6.5
10	18.6	26.4	24.4	7.8	18.0	23.4	22.8	5.4

Outstanding is the fact that an increase in the vegetation mass not only decreases the diurnal temperature oscillation (see

Δ 1300-0700), but also reduces the general level of warmth.

Under an abundant grass cover, the soil temperature is always lower.

Such a ratio of temperatures fully agrees with our concepts on the role of the grass cover as a thermal insulator and as a consumer of heat for evaporation.

The data presented in Figure 42, pertains to clear weather. As follows from Table 46 directly below, all the differences are naturally, smoothened out in overcast weather. The data in Table 46 were collected by A. I. Kameneva at Pushkino (Leningrad). For purposes of comparison, parallel data, pertaining to black fallow, is cited.

TABLE 46

Mean air and soil temperatures at 1300 hours amid standing wheat on clear and overcast days in July 1926.

Character of Soil Surface	Soil temperature		Air temperature	
	Surface 3-cm depth	10-cm depth	at vegetation height level	At 2-meter level

Clear Weather

Wheat	25.9	22.9	19.7	20.3	19.9
Black Fallow	32.6	26.4	20.2	20.7	

Overcast weather

Wheat	19.4	19.4	18.0	17.8	17.9
Black Fallow	21.2	20.5	18.7	18.0	

To characterize the effect of general climatic conditions,

we will analyze the geographic distribution on the European territory of the USSR on the basic factors that determine the microclimate of a wheat field, such as the radiation balance, the latent heat of evaporation, the amount of heat that goes into the warming of the soil and the air. In doing so, we will proceed from the geographic zonality of humidification under natural conditions, without considering the measures taken in the struggle with drought and dry winds, the effect of which measures will be analyzed additionally. The taking into account of natural zonality will allow for a better evaluation of the meteorological effectiveness of the indicated measures.

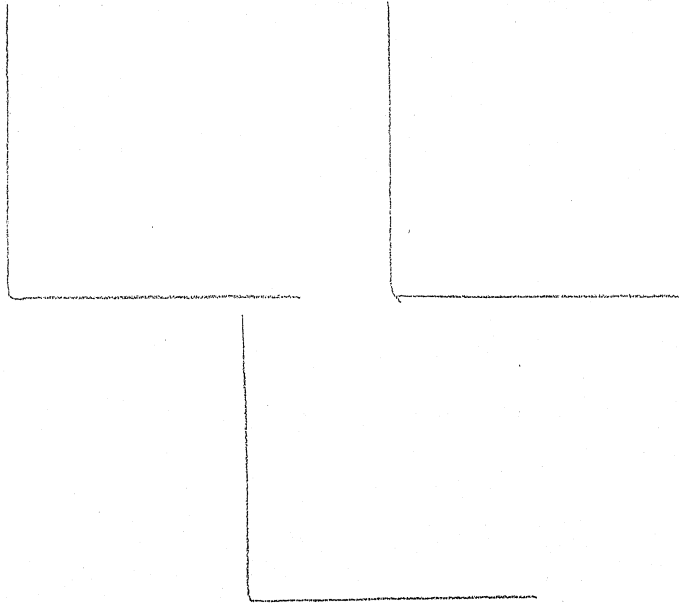


Figure 42. The thermal balance of a wheat field during the daylight hours, in its mean value, over the vegetation period. A- radiation balance (in calories), B-Amount of diurnal evaporation from summer wheat field and amount of heat consumed for that purpose (1-calories, 2-millimeters); V-the amount of heat that goes into the daylight warming of the soil and air (in calories/per cm^2).

In the above Figure 42, Section A shows the distribution diagram for the radiation balance of a wheat field during the daylight hours, in its mean value, for the vegetation period.

In order to compute the part of the radiation balance that goes into the direct warming of the soil and the air and determines the daylight temperature gradients, it is necessary to deduct the latent heat of evaporation from the radiation balance.

Section B of Figure 42 depicts the mean value of daylight evaporation over the period 1936-1940, as per the data collected by A. V. Protserov, Agro-meteorological Bureau GUGMS, and the corresponding expenditure of heat, taking into account the latent heat of evaporation.

It must be remembered that the evaporation data pertains to the parcels with natural humidification, and therefore, the said data and the values, that are tied in with them, are not to be considered as pertaining to the irrigated areas.

The amount of heat that goes directly into the warming of the soil and the air is depicted in Section V of Figure 42.

Without ascribing absolute values to the magnitudes presented diagrammatically in Figure 42, we can still utilize them for the comparative characterization of the conditions that go into the formation of the microclimate of a wheat field in the various climatic zones of the European territory of the USSR. Indeed the greater the thermal current, the greater, all other factors being equal, the temperature gradients. As to the absolute air humidity

gradients, they increase in relation to the intensity of evaporation. The greater the evaporation, the greater the humidity gradients.

The microclimatic characteristics during the nocturnal hours are determined by the intensity of effective radiation, which in the European territory of the USSR, increases from the Northwest to the southeast, principally, on account of the diminution in cloud formation occurring in the same direction. In addition radiation chilling in the northwest is retarded at the expense of the latent heat, which is liberated in the abundant condensation of dew. However, the effect of these two factors is somewhat overshadowed by the effect of the grass cover. The northwest grass cover of the greater denseness forms an almost continuous radiating surface, which, being insulated by the grass from the soil, acting during the night as a thermal storage battery, produces a reduction in temperature at the height of the plants. The temperature rises somewhat toward the surface of the soil. In the southeast, in the presence of a poorer grass cover under conditions of natural humidification, the minimum is either weakly pronounced or adapted to the soil surface.

The zonality of the phytoclimate is corroborated by observations conducted by the agro-meteorological stations, located near Leningrad, Saratov, and Rostov-on-the-Don.

Figure 43 gives a diagrammatic presentation of the vertical temperature variation in the lowest air layer and the top soil layer of a wheat field in the phase of inflorescence-ripening for the humidified forest zone and for the dry steppe zone (without irri-

gation or other measures for the raising of the soil dampness). The curves on the extreme right in each of the graphs characterize the variation of the maximum temperatures, usually timed to 1300-1400 hours; the curves in the middle - the variation of the mean diurnal temperatures; and the curves on the left - the variation of the minimum temperatures, observed before sunrise.

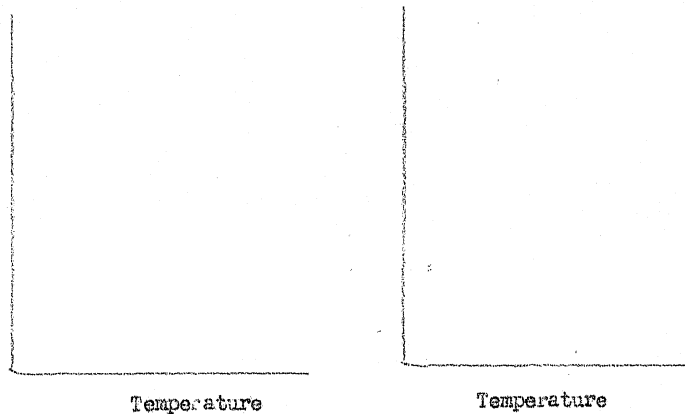


Figure 43. Diagram of vertical temperature variation in the lowest atmospheric layer and in the top soil layer of a wheat field in the phase inflorescence-ripening. A- in the humidified forest zone; B- in the dry steppe zone (without irrigation or other humidification measures); a- the height of the grass cover; b- soil surface temperatures: - mean diurnal; during period of maximum temperatures; during period of minimum temperatures.

Figure 43 well illustrates the zonal character of the microclimate of a wheat field against the background of the general temperature level and its diurnal range. It makes it possible to

arrive at the following conclusion with relation to fields, which are typical for the zones being compared, under conditions of natural humidification.

In the forest zone, the temperature of the air inside the grass cover during the entire diurnal period does not substantially differ ($\pm 1^{\circ}$) from the air temperature at the 2 meter height. The soil surface, to which the sun's rays practically do not penetrate, has a temperature approximating that of the air. During the period of maximum, somewhat higher temperatures are adapted to the air layers close to the soil, without, however, a sharply pronounced maximum at the very surface of the soil, while the lowest nocturnal temperatures are observed in the upper part of the grass cover.

In the dry steppe zone, in the non-irrigated fields, due to a poorer grass cover and to a greater solar altitude (not only to a lesser geographical latitude, but to an earlier ripening phase), a considerable part of the soil surface is illuminated by direct sunshine. As a result, the maximum temperature of the soil surface is higher than the temperature of the air at the 2 meter height by 20 degrees Centigrade and more. It is even higher than in black fallow, since the grass cover reduces turbulent exchange and consequently the heat emission into the upper air layers. In connection with this, the maximum air temperature in the lowest layer is by 2-3 degrees higher as compared to the temperature at the 2 meter height. The minimum temperature has an insignificant vertical variation (within the range of one degree), nevertheless the minimum temperature value is adapted to the near-the-surface-layers. The mean diurnal temperature in the lowest

air layer rises somewhat, approximately by 1 degree.

Taking into account the aforesaid, it can be acknowledged that, in comparing the agro-climatic factors in the growing of wheat in the forest and the steppe zones on the basis of the mean diurnal temperatures the micro-climatic correction is not particularly significant. However, in the evaluation of the daylight temperatures, adapted to the most vital hours in the active life of plants, the microclimatic correction assumes a substantial value.

Such are the typical aspects of the microclimate of a wheat field in the humidified and the dry zone. It does not follow from this that, in individual cases, even in the northwest, sharp vertical temperature gradients may not occur in the midst of a sparse grass cover, particularly during the hot and dry summer. On the other hand, the measures undertaken in the dry zone for the increase of the soil moisture reserve, such as snow retention, increase the losses of heat to evaporation, reducing thereby the direct flow of heat into the soil and into the air, and, consequently, the temperature gradients.

Irrigation exerts a decisive effect upon microclimate. The effect of irrigation upon the phytoclimate of wheat was studied in 1934-1937 in the Saratov Oblast, in the Yershovo irrigated area of the Institute for the Grain Economy of the Southeast. The basic object of study was summer wheat ("Melyanopus 069") under two-time flood irrigation treatment. L. A. Golubera summarized the results of these experiments in a series of articles. The effect of irrigation upon the phytoclimate of a wheat

field is particularly great, in the presence of the full development of the grass cover in the days immediately following irrigation.

As an example, we will analyze the effect of irrigation upon the thermal cycle of the soil (Table 47).

TABLE 47

Temperatures of the top layers of the soil at 1300 hours (1934-1935)

Phase of growth	Soil at the depth of					
	3 centimeters			10 centimeters		
	non-irrigated	irrigated	Δ	non-irrigated	irrigated	Δ

After first irrigation

Fasciculation-spike formation	28.5	24.5	4.0	21.6	19.8	1.8
Spike maturity	26.4	26.0	0.4	20.8	20.0	0.8

After second irrigation

Inflorescence-grain ripening	31.7	21.9	9.8	23.9	20.8	3.1
Finish Ripening	27.7	24.6	3.1	22.3	21.4	0.9

As can be seen from the Table above, the mean temperature of the soil at the depth of 3 centimeters, in an irrigated area, was almost 10 degrees Centigrade lower, as compared to a

non-irrigated area. The temperature and relative humidity of the air in the grass cover of irrigated and non-irrigated wheat, and also at the height of 200 centimeters (observation booth) are cited in table 48.

The maximum differentials in both the thermal cycle and humidity are adapted to the inflorescence grain-ripening phase in the growth of wheat. [Table 48 on next page]

It is important to note that, during the semi-arid and dry wind days, the contrast between the irrigated and non-irrigated fields becomes even greater than usual. Thus, during the inflorescence-grain ripening phase in the growth of wheat, on dry wind days, in the irrigated areas, at the height of the working leaf, the temperatures of the air is lower by 3 degrees Centigrade, and relative humidity is by 15 percent higher than in the non-irrigated areas.

Due to the fact that, at the height of 2 meters and above, the air temperature is determined not only by the local cycle, but also by the heat inflow from the adjacent non-irrigated areas (local advection), the temperature of the air in the indicated layer is, in a number of cases, higher than the temperature in the midst of the grass cover. In other words, over the irrigated areas, there occurs frequently even in daylight a temperature inversion and, consequently, a diminished turbulence. This phenomenon is important - it somewhat retards the further increase in evaporation. A poor aeration of the irrigation fields, not only in the soil, but also in the midst of the grass cover, is also tied in with it.

TABLE 48

Temperature and relative humidity of the air at 1300 hours

Phase of Growth	Air at the height of						200 centimeters
	10 centimeters			the effective leaf			
	Non-irri- gated	irri- gated	△	non-irri- gated	irri- gated	△	

Temperature of the air

After first irrigation

Fasciculation-tube formation	24.2	22.4	1.8	23.4	22.2	1.2	22.8
Stem formation spike formation	27.2	25.4	1.8	25.4	25.4	1.0	25.8

After second irrigation

Inflorescence- grain ripening	30.6	26.3	4.3	29.1	27.0	2.1	28.0
Finish-maturing	28.9	27.4	1.5	27.2	26.4	0.8	26.1

Relative humidity of the air (in percent)

After first irrigation

Fasciculation tube formation	48	62	14	48	59	11	36
Stem formation spike formation	45	56	11	40	49	9	34

After second irrigation

Inflorescence- Grain ripening	34	62	28	34	48	14	33
Finish-maturing	47	55	8	46	51	5	40

Chapter 9

THE COMPARATIVE ANALYSIS OF SOME FEATURES IN THE
MICROCLIMATE OF INDIVIDUAL AGRICULTURAL CROPS

The comparative analysis of the microclimate of various agricultural crops is very difficult. Even in those cases when the observations of the crops being studied are conducted simultaneously and by the use of the same methods, it is difficult to make sure that the derived correlations of the elements of the phytoclimate are really determined by the species and variety-composition of the crops, and not tied in with an accidental sparseness or abundance of the grass cover, or peculiarities of the soil and agricultural technique.

Table 49 (as per N. A. Kurmangalin) gives a visual illustration of the manner in which one of the phytoclimate elements - the temperature of the soil, varies under the effect of agricultural technique and diversity of soil.

TABLE 49

The soil temperature in a potato field at the
depth of 15 centimeters, Khibiny, 1934

Character of the	VI		VII		VIII		IX		
Top Soil	0200	0300	0100	0200	0300	0100	0200	0300	0100
Mineralized, ridges	8.0	13.0	12.4	18.4	17.2	14.9	11.3	10.9	9.9
Pete Ridges	7.7	9.9	11.9	15.9	15.5	14.4	11.4	11.1	9.9
Level field	6.3	9.0	9.7	14.8	14.7	13.8	11.1	11.0	11.0

With the development of the plant and the consolidation of the grass cover, the microclimate of the field must change. The microclimate of different varieties of the same plant species, differing from each other only by the size and disposition of their leaf blades, is different.

At the same time, the microclimate of plants belonging to different species and even to different genera, as, for instance, oats and wheat, may be homogeneous, when their respective vegetation mass, the character of their foliation, and the transpiration factor are homogeneous.

As was already indicated, all these characteristics were, up to a recent date, not taken into account in the microclimate observation procedure and, therefore, the data that follows basically characterizes the possible differences only, without guaranteeing their typicalness, particularly quantitatively.

The comparative analysis of the microclimate of agricultural crops, due to the difficulties involved, is usually limited to the consideration of its individual components.

Thus, during the summer of 1937, the Agro-hydro-meteorological Institute conducted comparative minimum temperature observations in the case of seven agricultural crops (Table 50). [Table 50 on next page].

Table 50 above cites the differentials of the mean 10 day period minimum temperatures, readings taken by exposed thermometers installed at the heights of the individual crop plants and on the

TABLE 50

Minimum temperature differentials in the surface of denuded soil at the meteorological platform and at the height of the crop growth - Ovtstino, 1937

Crops	VII	VIII		IX		
	0300	0100	0200	0300	0100	0200
Flax	0.1	-0.2	0.2	0.0	-0.2	-0.8
Wheat	0.4	-0.1	0.0	-0.3	-0.2	-0.8
Potatoes	0.7	0.3	0.4	0.6	-0.2	-0.6
Tomatoes	0.0	-0.3	-0.2	-0.4	-0.4	-0.9
Sunflower seeds	0.0	-0.4	-0.2	-0.8	-0.7	-
Clover						
Oats	0.6	0.2	0.4	-	-	-
Proso millet	-	-	-	-	-	1.3

surface of the denuded soil. First of all, it is necessary to remember that the mean minimum temperature at the grass cover height differs only insignificantly (usually, by less than 1 degree) from the minimum temperature at the surface of the denuded soil of the meteorological platform. Most frequently, the minimum at the grass cover height is higher than the minimum temperature of the soil. Only in potatoes and in the mixture clover-oats lower minimums are observed.

The differentials between the minimum temperatures over individual crops mostly do not exceed one degree Centigrade.

For the group of crops (flax, wheat, tomatoes) the minimum temperature differentials do not reach 0.5 of a degree. The corroboration of this by following observations will have great methodical value, since it will simplify considerably the study of the early frosts in the midst of agricultural crops.

Interesting data was collected by observations of soil temperature at the depth of 10 centimeters, conducted at Ovtino in July and August of 1937, and involving 13 agricultural crops. The peculiarities of the thermal cycle of the soil make it possible to divide all the indicated crops into 3 groups. (Table 51 on next page].

At the setting in of the minimum temperatures, which was observed at 0600-0700 hours, the temperature differentials were smoothened. Thus, on 5 August they were less than 2 degrees, with the lowest temperature observed under the beets, clover, peas and potatoes.

As can be seen from Table 51 (according to A. I. Andreyeva), the minimum diurnal range (1.4 - 2.2 degrees), as well as the lowest maximum (16.4 - 17.2 degrees) are adapted to the crops with the greatest green mass (peas, clover, millet) and to meadows, which in addition to its considerable green mass is distinct by the presence of sod.

Group II, with a diurnal temperature range of 2.3 - 3.7 degrees and a maximum temperature of 17.6 - 18.6 degrees, contains crops with a lower green mass, such as the aftermath of a meadow, beets, flax.

TABLE 51

Temperature of the soil at the depth of 10 centimeters on 5 August 1937

Temperature	Group I					Group II				Group III				
	Meadow	Podol'sk Clover	Peas	Proso millet	Pskov Clover	Meadow Aftermath	Beets	Flax	Sunflower	Wheat	Potatoes	Meteorological platform grass cover	Clover aftermath	Fallow at meteorological platform
Maximum	16.6	16.4	16.4	17.0	17.2	18.4	17.6	18.4	18.6	19.8	19.2	20.2	20.6	20.4
Minimum	15.2	14.8	14.8	15.0	15.0	16.1	14.7	15.4	14.9	15.8	14.8	15.6	15.6	15.1
Range	1.4	1.8	1.8	2.0	2.2	2.3	2.9	3.0	3.7	4.0	4.4	4.6	5.0	5.3

Note: During the hours of the maximum temperatures, the differentials are the greatest.

Group III consisted of wheat, potatoes, the grass cover of meteorological platform, the aftermath of a clover field, and the denuded meteorological platform (fallow)

The soil temperature under the crops of this group is distinguished by a considerable wider range (4.0 - 5.3 degrees). The maximum temperature under individual cultures oscillates from 19.2 to 20.6 degrees Centigrade. The grass cover at the meteorological

platform, free from sod and trimmed frequently, has a temperature, which is substantially different from the temperature of a natural meadow (by 4 degrees of maximum temperature).

Some interesting temperature data relating to the soil surface is furnished by M. N. Kopachevskaya for a beet crop (non-irrigated) during the first and second year of its growth.

Figure 44 is a graphic representation of the maximum temperature cycle of the surface of the soil by pentades (five-day periods) under non-irrigated beet crop during the first and the second year of growth and the temperature cycle of black fallow. The temperature cycle under beets in the first year of vegetation at first approximates that of black fallow. But from the middle of June, with the development of the beet leaves and the formation of a dense grass cover, the temperature of the soil is reduced considerably and approximates the temperature of the air at the height of 2 meters. The temperature cycle of the soil surface under the beets in its second year of growth (transplanted for seed) is different. Regardless of the developing stems and abundant leafing, attaining its maximum at the time of mass inflorescence, the surface of the soil is not shaded, by virtue of which fact its temperature is close to the temperature of black fallow during the entire summer, and at times even exceeds it. The cited example is of particular interest, since it points out the extent to which the microclimate of a crop varies with the process of its growth and with the change in the character of its leafing.

Temperature in degrees

Stem development
of transplantings

Browning of the
tubers

Mass inflor-
escence

Figure 44

Pentades

Maximum temperature cycle of the soil surface, by pentades.
Ivanovsk Selective-Experimental Station, 1938.

1 - under beets; 2 - under transplantings. 3 - black fallow;
4 - in the air (psychrometer observation booth).

In line with the general regularities in the formation of microclimatic characteristics, the latter become more pronounced with the transition to the southern continental zones. The microclimate of southern crops, most analyzed and studied, is the microclimate of cotton and lucerne (alfalfa).

L. I. Babushkin, I. G. Sabinina, and A. V. Shtyrev, who conducted their research at the Boz-Su Agro-meteorological Sta-

tion (near Tashkent), furnish the characteristics of the types of vertical temperature gradients of the air over cotton fields and alfalfa.

By the character of the temperature variation with height, within the layer from 1 to 200 centimeters (observations made at 1, 25, 50, 100, and 200 centimeters), they segregated 4 types of gradients, out of which we will mention two: the temperature inversion and the superadiabatic gradient in the entire layer. The recurrence of various values of the temperature differentials for the layer 25-100 centimeters (in the case of the temperature inversion), and for the layer up to 200 centimeters (in the case of the superadiabatic gradients), as observed at 1300 hours over cotton and alfalfa, are given in Table 52.

TABLE 52

The recurrence of various gradations of temperature differentials in the 25-100 centimeter air layer
(in percent of total number of observations)

	Temperature differentials					Total
	0.1-0.5	0.6-1.5	1.6-3.5	3.6-5.5	5.6-7.5	
			Cotton			
Inversion	5	3	0	0	1	9
Superadiabatic Gradient	11	27	16	1	0	55
			Alfalfa			
Inversion	7	9	20	10	2	48
Superadiabatic Gradient	8	3	1	0	0	12

Both, cotton and alfalfa are irrigation crops, but the character of their grass covers are substantially different, which is reflected in their thermal cycles. In the case of cotton, a temperature inversion during the daylight hours was observed only in 9 percent of all observations, and the temperature differential was not in excess of 1.5 degrees Centigrade, while, in the case of alfalfa, on the contrary, almost in half of all observations, there was a temperature inversion, with differentials above 1.5 degrees predominating. In individual cases, the temperature differentials attained the value of even more than 5 degrees Centigrade. Superadiabatic gradients predominated over the cotton fields.

In nearing the surface of the soil, the temperature gradient, naturally, is increased. Table 53 depicts the character of the vertical temperature variations within the 1-50 centimeter air layer over a cotton field for 2 periods in its growth: up to buttonization, when the height of the cotton is less than 50 centimeters (up to 15 June), and beginning with inflorescence (after 15 July) to the end of its growth, with the height of the cotton over 50 centimeters (as observed at 1300 hours)

TABLE 53

Recurrence of various gradations of temperature differentials in the 1-50 centimeter air layer (in percent of the total number of observations)

Period	Temperature differential							
	1	5	19	26	24	18	6	1
Up to 15 June	1	5	19	26	24	18	6	1
After 15 July	13	31 [?]	37	13	3	2	1	0

With progress in the growth of cotton, the vertical temperature variation is less pronounced, and if, during the first period of growth, temperature differentials ranging from 3.6 to 7.5 degrees were predominant, their range, during the second period, was only from 0.6 to 3.5 degrees Centigrade.

For the additional characterization of the microclimate of alfalfa, data illustrating the effect of mowings upon the vertical temperature distribution of the air at 1300 hours is depicted in Table 54.

[See Table 54 on following page]

As seen from the above table, after mowing down the basic mass of the alfalfa, the thermal cycle of the parts of the plants remaining above the surface is changed abruptly (approximately, by 10 degrees Centigrade). Of course, the thermacycle of the roots is changed to no less a degree.

TABLE 54

The effect of alfalfa mowings upon the vertical
temperature distribution

Alfalfa	Height in meters		
	0	0.5	2.0
Before mowing	29.3	29.0	33.1
After mowing	40.1	34.2	34.0

In summarizing the above analyses, we come to the conclusion about the great diversity of the microclimate of agricultural crops.

These microclimatic differences are of such an order that, in a number of cases, they can promote the early ripening of plants, in other cases, on the contrary, retard it, it can increase or decrease their frost endurance, it can safeguard the plants against fungus diseases or increase the predisposition (to same.) All this proves the practical importance of controlling the microclimate (see Chapter 17).

Of self-contained interest is the climate of the peat marshes, which in some individual zones occupy vast areas. Numerous investigations have shown that evaporation, and, consequently, loss of heat to evaporation, from a natural moss swamp is lower than evaporation from mineralized parcels covered with abundant herbaceous vegetation.

The basic factor, determining the climatic features of marshland, is the thermal characteristic of peat - its low heat conduction. The thermal characteristics of peat run in relation to its humidification. The thermo-insulating effect of peat is diminished as its humidification increases, with a simultaneous increase in its thermal capacity. Thus, the climate-forming role of peat changes in relation to its moisture content. The maximum variation of the latter occurs during the summer, in relation to it being a dry or a rainy year, as well as in space, in relation to it being a drained or a non-drained area, the upper layer of moss removed or left intact. During the rest of the year, when atmospheric precipitation is in excess of evaporation, parcels, that undergo natural or artificial draining, are humidified, and, by its physical properties, again approximate the peat in the non-drained areas.

This variability in the moisture content of peat is partially the cause of the contradiction and uncertainty in the conclusions of some individual investigators of the climatic characteristics of peat bogs, particularly of their thermal cycle. The predominant adaptation to lowlands, at times even to drainless hollows, into which cold air flows off and becomes stagnant, also has its effect upon the climate of the peat bogs, usually leading to accelerated occurrence of frost.

Figure 45 is a comparative graphic presentation of the annual temperature variation in mineralized soil and in peat soil, drained and non-drained.

Figure 45. Annual variation in the temperature of the soil:
 mineralized (1), peat, drained (2), peat, non-drained (3).
 A-depth 0.4 meters; B- depth 0.8 meters.

The graph shows, first of all that, with depth, the difference in the thermal cycle, as between peat and mineralized soil, becomes greater in the summer as well as in the winter. This proves that the difference in the thermal cycle is determined by the properties of the soil itself, and not by other factors outside the soil (such as, for instance, the amount of absorbed radiation or the amount of heat lost to evaporation), for in the latter case the differences in the thermal cycles would be smoothened instead of accelerated. The depth of frost penetration in peat bogs is not as great as that in mineralized soils, but in the thawing out of frozen soil in peat bogs, particularly drained ones, requires considerably more time. In the environs of Leningrad, in peat ground, there were cases of locating frozen spots at a depth ranging from 30 to 75 centimeters up to the second half of July. In the drained swamps of the Kola Peninsula

even a zone of perma-frost was discovered, it obviously being the case that the formation of the perma-frost zone was induced specifically by the draining of the swamps with the accompanying reduction in the heat conduction of the ground during the dry summer season, i.e. exactly at the time of the thermal flow into the soil.

Data collected by S. G. Parkhomenko confirms that the persistence of frozen zones in the soil is specifically due to the dryness of the soil, by bringing out the fact of the disappearance of frozen passes after a rainy summer.

The diurnal temperature variation in peat soil is leveled off at the depth of 20 centimeters, but on the surface of dry peat soil it is very great, having a greater value than on the surface of mineralized soil.

The removal of the top layer of moss, not as yet disintegrated (the combing off), is conducive to a reduction in the diurnal temperature range of the soil surface layers.

[See Table on following page]

As can be seen from Table 55 (data collected by the Agro-hydro-meteorological Institute), the temperature change from 0200-0300 hours to 1200-1300 hours, on the surface of peat, with the top layer of moss removed, attained a value of 17 degrees Centigrade, while on the surface of peat, with the top layer of moss left intact, it was only 14 degrees Centigrade. The same differential stands for the near-the-soil air layer.

TABLE 55

Temperature of the soil and near-the-soil air layer in
the case of a peat bog, Murmansk (Sovkhoz Rosta), 22 June 1934

Peat	Soil (depth in centimeters)			Air height 2 centimeters
	15	5	0	

0200 - 0300 hours

Top moss

removed	6.5	8.4	8.0	9.6
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Top moss

intact	6.6	8.2	10.5	10.4
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1200 - 1300 hours

Top moss

removed	6.3	12.9	25.0	24.3
---------	-----	------	------	------

Top moss

intact	6.7	10.1	24.1	22.6
--------	-----	------	------	------

Observations conducted in the Ramensk swamp over the period 1914-1917 showed that the temperature of the soil surface in June was oscillating from minus 10 degrees Centigrade (absolute minimum) to plus 51 degrees Centigrade (absolute maximum), the respective oscillations in July being from minus 7 to plus 60 degrees Centigrade.

A. D. Dubakh furnished proof that even in the area Moscow-Kalinin-Novgorod, not to mention areas further to the North, in

slightly drained swamps stretching over tens of thousands of hectares, freezeovers occur on the surface of the soil during all summer months, not sporadically, but regularly.

CHAPTER 10

THE MICROCLIMATE OF SLOPES

Even under conditions of the plain which are predominant in the European territory of the USSR, variously oriented [exposed] slopes comprise a considerable percent of the general area, by virtue of which the microclimate of slopes is of great practical interest.

The microclimate of slopes with a northern as well as with a southern exposure has a great effect upon the climatic boundaries of plants. Due to the specific characteristics of the microclimate of slopes, plants, including the cultivated varieties, may reach out far beyond the boundaries of their continuous propagation.

The exposure and the steepness of the slope determine the local character of the inflow of solar radiation (see printed pages 21-22). But the characteristics of slopes are not limited to this only. As a rule, slopes are more readily drained than level places. The snow on slopes exposed to the south melts away earlier, due to the greater inflow of solar radiation. The latter is conducive to a rapid evaporation of the moisture detained in the soil, as a result of which the grass cover on the southern slopes becomes somewhat scarce. The diminution in the moisture content of the soil is pronounced particularly sharply on southern slopes, where the runoff of water is stronger due to the small extent of sogginess. Southern slopes are covered predominantly with attenuated soil, which cannot retain water.

In the final analysis, the microclimate of slopes is

determined not as much by the difference in solar radiation income as by the secondary factors, such as the moisture content of the soil and the capacity of the grass cover.

The wind exerts a strong effect upon the microclimate of slopes by intensifying the emission of heat from the soil into the air.

The complexity of the factors, which determine the microclimate of slopes, make the study of the regularities in its formation difficult.

As an example of the sharp effect of slope exposure upon the temperature of the soil, as an element of microclimate, the results of observations by the students of climatology of the Leningrad State University are cited. The observations were made in clear calm weather near Leningrad, on a hill with a 20-22 degree slope (Table 56). The northern slope was covered with an abundant grass cover with dense sod, the soil being damp. The grass cover on the southern slope is not in close formation, the upper layers of the attenuated soil being dry.

[see Table on following page]

The considerable difference in the temperature of the denuded soil is determined not only by the difference in solar radiation income, but also by the intensive evaporation from the damp soil, of the northern slope.

With a lower degree of steepness (10 - 15 degrees), in the absence of substantial differences in the soil itself and its

TABLE 56

Temperature differentials in the soil as between the southern and northern slopes at the depth of 10 centimeters (the southern slope is warmer than the northern) Sabbino, 17 July 1938

Soil	Hours						
	1000	1100	1200	1300	1400	1500	1600
Denuded							
(on 14 August	8.4	9.8	11.8	14.0	16.1	15.6	15.7
Grass covered	3.2	3.4	4.3	5.4	6.2	6.8	7.4

cover, and particularly in the presence of a south wind, the temperatures of the northern and southern slopes are practically the same. As an example, the results of observations of the Agro-hydro-meteorological Institute also in the environs Lenin-grad, can be cited (Table 57).

[see Table on following page]

The relative drop in the temperature of the denuded southern slope (fallow) as compared to the northern was due to the fact that no wind was blowing over the southern slope during the period of observation. The soil somewhat shielded from the wind by meadow vegetation, against the background of a lower (than in fallow soil) temperature, was still warmer at the southern slope as compared with the northern slope.

TABLE 57

Temperature of the soil in the southern and northern slopes during clear weather. Ovrino, 13 July 1936

Character of	Southern slope		Northern slope	
Soil Surface	Depth - in centimeters			
	5	15	5	15
1100 - 1230 hours				
Fallow	24.5	19.8	25.4	20.0
Meadow	21.4	18.6	19.2	17.1
0233 - 0425 hours				
Fallow	15.9	17.8	15.7	18.2
Meadow	17.4	17.7	15.8	16.7

An example of the effect of slope exposure upon the temperature of the air is given in Table 58.

As seen from Table 58, the slope exposure has a substantial effect only upon the near-the-soil air layer. As was to be expected, the temperature differentials are leveled off during overcast weather. The minimum temperature differentials were due not to slope exposure but to the fact that the southern slope was facing the sea, which is warm during the night, while the northern slope opened up into a valley that was shielded from the warm effect of the sea.

TABLE 58

Mean temperature of the air over a 9-day period
over the southern and northern denuded slopes
(steepness about 20 degrees) in the Batumi Botanical Garden, Sept. - Oct. 1935

Slope	Depth - in centimeters				
	5	25	50	100	150
1300 hours, clear					
Southern	28.0	25.2	23.8	22.6	22.0
Northern	23.8	22.6	22.2	22.0	21.8
Differential	4.2	2.6	1.6	0.6	0.2
1300 hours, overcast					
Southern	24.4	23.0	22.5	21.8	21.6
Northern	23.4	22.4	22.1	21.9	21.8
Differential	1.0	0.6	0.4	-0.1	-0.2
Before sunrise, clear					
Southern	17.7	18.8	19.3	19.6	19.8
Northern	16.8	18.0	18.9	19.1	19.4
Differential	0.9	0.8	0.4	0.5	0.4
Before sunrise, overcast					
Southern	20.1	20.4	20.7	20.9	21.0
Northern	19.7	20.0	20.3	20.4	20.6
Differential	0.4	0.4	0.4	0.5	0.4

Note: During the day, a westerly wind predominated,
changing to an easterly during the night.

All other conditions being equal, the minimum temperature at the southern slope is approximately the same as at the northern one.

Table 59 cites the diurnal temperature variation of the air at the height of 28 and 150 centimeters over the denuded soil of a southern and northern slope (see text and Table 56). While, during the day, at the height of 25 centimeters, the differentials attained the value of 5 degrees, at 0400 and 2000 hours, they were only 0.6 - 0.7 of a degree. At the height of 150 centimeters, the temperature differentials are small both, during the day and the evening.

TABLE 59

Temperature of the air over the southern and northern slope, Sabbino, 16 July 1938

Slope	Hours									
	0400	0600	0800	1000	1200	1400	1600	1800	2000	
At the height of 25 centimeters										
Southern	14.9	19.7	28.4	32.0	32.6	35.8	32.6	26.5	22.4	
Northern	14.3	21.8	27.8	31.4	31.2	30.8	28.6	25.0	21.7	
Differential	0.6	-2.1	0.6	0.6	1.4	5.0	4.0	1.5	0.7	
At the height of 150 centimeters										
Southern	14.9	19.3	26.7	30.4	30.4	30.9	30.1	25.4	22.0	
Northern	14.6	20.7	27.1	29.9	30.5	30.6	29.6	24.1	21.9	
Differential	0.3	-1.4	-0.4	0.5	-0.1	0.3	0.5	1.3	0.1	

However, in the presence of uniform minimum temperatures, the slopes differ by the degree of frost hazard. The most frost-bitten are the eastern and southeastern slopes, due to the sharp rise in temperature during the morning hours, followed by a rapid defrosting of vegetation tissues resulting in damage to the latter.

The isopleths of the air temperature on eastern and western slopes, as per observation by V. A. Smirnov, are represented graphically in Figure 46.

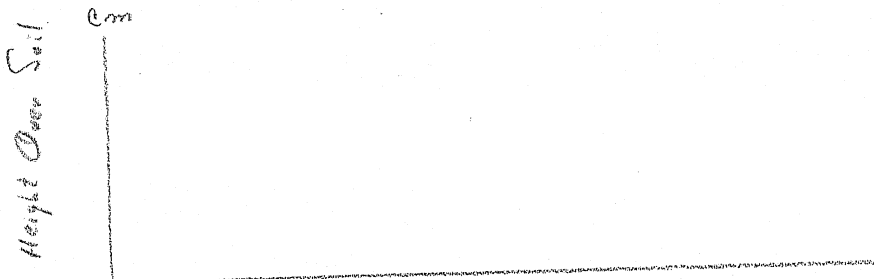


Figure 46. Temperature isopleths in the near-the-soil air layers over the various exposure slopes of Abkhazia -Sovkhoz "Il'yich", 1935- A- eastern slope; B-western slope.

During the morning hours of 23 February 1935, in the area of Sukhumi, over the eastern slope, already at 0820 hours, the insolation type of vertical temperature distribution is observed:- at the soil surface 9 degrees, at the height of 1.5 meters 6 degrees. While over the western slope, at this time, there is still a temperature inversion, and the temperature at the soil is 3 degrees Centigrade. By 0920 hours, over the eastern slope, the temperature at the soil attains a value of 12 degrees Centigrade, while over the western slope, it is only 5 degrees. This example well illustrates the rapid morning rise in temperature

over the eastern slope as compared to the western.

In exerting a direct effect upon the thermal cycle of the air and the soil, the exposure of slopes also controls air humidity.

As can be deducted from Table 60, substantial humidity differentials are also adapted to the near-the-soil air layers only.

TABLE 60

Relative humidity (in percent) in Batumi Botanical Garden, September-October 1935, 1300 hours, clear

Slope	Height - in centimeters				
	5	25	50	100	150
Southern	54	63	64	69	70
Northern	70	71	72	72	72
Differ- ential	-16	-8	-8	-3	-2

The effect of slope exposure on microclimate is a good example of reciprocity of physico-geographical factors. The character of the microclimate of slopes determines their soil and vegetation covers, the latter, in turn, aggravating the microclimatic differentials. From this, the conclusion can be drawn that the advantages and disadvantages of the microclimate of slopes may be either intensified or attenuated by varying the vegetation and the soil.

In order to increase the thermal advantages of southern slopes under the surrounding conditions of northern country it is necessary to select parcels with mild wind velocities, or to shield them from the wind.

The deduction on the predominant effect the exposure of slopes has upon the microclimate of the near-the-soil air layers and upon the soil is not in contradiction with the fact, that the exposure of slopes (for instance, in Central Asia) has a determining effect upon ligneous vegetation. The normal development of hazel nut and maple forests on the northern slopes of Gissarsk Mountain Range, and their complete absence on the southern slopes, is determined by the ample moisture supply originating with the cold season precipitation and accumulating in the mighty soil layer (up to 2 meters thick), while the attenuated soil of the southern slopes is not capable of retaining this moisture.

It follows, then, that the presence of forests on the northern slopes is, in this case, determined by the climate of the soil and its conginguous air layers.

CHAPTER II

THE SNOW COVER AND ITS MICROCLIMATE

The character of the disposition of the snow cover runs in direct relation to local conditions. It is affected not only by ligneous vegetation, hillocks and ravines, but also by remnants of herbaceous vegetation, ridges, and furrows in plowed fields. Snow is easily blown off from parcels of land exposed to great wind velocities, and, inversely, it becomes accumulated in places where there is calm, the decisive factor being the velocity of the wind at the very snow surface. This latter circumstance is of great practical importance, since it shows that in order to control the snow cover, it is only necessary to change the velocity of the wind in the very lowest layers of the air.

Let us consider the distribution of the snow cover on several elementary forms of the terrestrial relief in the presence of a prevailing definite direction of the wind. Figure 47 shows diagrammatically the distribution of snow accumulation at a vertical wall (escarpment), in a trench with shallow slopes, at a narrow ridge with steep and shallow slopes, and at an isolated small elevation (according to G. D. Rikhter).

Figure 47. Diagrams of snow accumulation on elementary forms of terrestrial relief.

At the base of the windward steep wall, due to vorticity, a deflation spout is formed. The snow is being thrown from the spout toward the wind and is forming into drifts. Over the edge of the escarpment, the capacity of the snow is reduced by blowoff

(Figure 47,A). At the edge of a leeward escarpment (Figure 47,B), a beak-shaped snow visor, having at times a downward bend, is formed. A drift, swept up by vortices, is formed at the base of the escarpment.

A trench with shallow slopes (a railroad ditch, a brook trough, a rivulet), at its leeward slope, has a snow drift with curled visor formation, while the windward slope has no snow. The accumulation of the drift at the leeward slope may overfill the ditch completely (Figure 47, V). However, very deep troughs may be considered practically non-fillable, since the eddies forming in them keep throwing the snow out. The depth of such non-fillable troughs varies with different zones.

The snow is blown off from the shallow windward slope of the narrow ridge (Figure 47, G), while at the top of the leeward steep slope, the snow forms a cornice, and at the base of it, it is formed into drifts.

An individual small elevation (a hillock, a mound, a haystack) has, from the leeward side, a deflation spout, behind which there is a flat tongue-shaped drift. A deflation spout usually forms also from the windward side (Figure 47,D).

The snow cover mostly smoothes the irregularities in the relief, including the terrace-like ledge at the valley slope, small troughs, and the like.

For a quantitative estimate of the snow cover capacity on different relief formations, the results of observations by V. P. Mosolov on 3 slopes in the Kamsk-Ust'yinsk area of the Tatar ASSR in 1940 (Table 61).

The upper part of the slope has the least amount of snow cover, while at the lower part of the slope and in the valley, its capacity is greater.

TABLE 61

Height of the Snow cover on 3 slopes
(in centimeters)

Location	Slopes		
	1	2	3
Watershed	31	25	23
Upper Part	24	17	16
Middle Part	22	30	16
Lower Part	31	33	18
Valley	60	94	28

Climatological manuals, in describing the disposition of the snow cover, use the terms "shielded" and "exposed" parcels of land.

A shielded land parcel is a parcel protected from the wind - in a garden, a populated spot, in a glade. An exposed parcel is a parcel of land accessible to the wind. Table 62, as an example, cites the capacity of the snow cover at 2 such spots, located in direct proximity, but differing as to wind accessibility. The mean value of the snow cover over the shielded parcel was almost twice that over the exposed one.

TABLE 62

Height of the snow cover as of the 15th of each month
(in centimeters)

Parcel	XI	XII	I	II	III	IV
Shielded	4	16	31	46	50	19
Exposed	3	10	16	26	26	3

Excessive accumulation of snow and snow drifts are derogatory, particularly to transportation. The combating of snow drifts is based on the principle of holding the snow to the side of the road with the aid of various obstructions (fences, shields, live hedges). The most widespread method is by continuous shielding made of wooden planks with gaps in between. They are durable, and they attract great snow drifts on both sides. A flat drift is accumulated from the windward side of the shield. In the direct proximity of the shield there is no snow accumulation (Figure 47a)

Figure 47a. Diagram of snow accumulation at a continuous shield.

The maximum snow accumulation occurs at the leeward side of the shield, with the drift usually forming a beak-shaped ridge. When the snow ridge on the leeward side of the shield attains three quarters or the entire height of the shield, the shield dugout begins

to fill, and the snow will then pass over the obstacle that was put in its way. To avoid this, during great accumulations of snow, the shields are shifted onto the top of the snowdrift.

The effect of live hedges is similar to the effect of echelon formations or forest belts. By accumulating snow in themselves, they protect the railroads, highways, and other roads.

The effect of forests and field shielding belts on the capacity of the snow cover is discussed in the corresponding chapters. In Chapter 17, the meteorological effectiveness of special snow-detaining measures is discussed. Here we will center our attention on the climate-forming role of the snow cover, on its effect upon the thermal and radiation cycles of the lowest atmospheric layers and, particularly the soil.

The great reflecting capacity of the snow cover was discussed in Chapter 1. This snow characteristic plays an important part in its thermal cycle, and retards its warming and melting. This circumstance specifically accounts for the fact that, in the process of snow melting, the advective heat may compete with radiation heat.

The increase in diffused radiation at the expense of reflection from the snow cover results in an increased radiation balance of the adjacent surfaces, that are free from snow, thereby increasing further the contrast by heating them. The melting of snow is more intensive at a tree trunk, at the stems of the grass cover, with the formation in the snow cover of peculiar "funnels of melting".

The snow cover not only reflects solar radiation, but also allows its permeation. The transparency of snow varies in relation to its structure, air impregnations, their form and size, the degree of water filling. There are reasons to assume that some effect upon its transparency is exerted by the temperature of the snow, since the content of water vapors in the air spaces of the snow is related to temperature.

The transparency factor P of snow is determined by P. P. Kuz'min by the formula $I_z = I_0 P^z$, where I_0 is the intensity of radiation entered into the snow cover through its upper boundary; I_z is the intensity of radiation permeated into the snow to below level z . According to N. M. Kalitin, the transparency factor for dry snow is approximately 0.90 and for wet snow - 0.65.

Table 63 depicts the layer-wise absorption of total solar radiation at various values of the transparency factor (as per P. P. Kuz'min). The first 5-centimeter layer of snow, in relation to the transparency factor, absorbs from 34 to 88 percent of the solar energy that entered the snow. The 10-centimeter layer absorbs from 56 to 98 percent. Under the most favorable conditions, radiation is absorbed down to the 60-centimeter layer. Hence, with the snow cover heavier than 60 centimeters, all the snow layers below that level and the soil underneath the snow are practically insulated from the sun.

[see Table 63 on following page]

Ice has a much greater factor of transparency, 0.96-0.98. Thus, through ice that is 7 centimeters thick, more than

TABLE 63

Layer-wise absorption of total solar radiation, at various values of P, in percent of radiation that entered the snow.

P	Δz								
	60 cm								
0.92	34.4	22.4	14.9	9.7	10.7	4.6	2.0	0.9	0.7
0.90	41.0	24.2	14.2	8.5	7.9	2.8	1.0	0.3	0.1
0.88	47.2	24.9	13.2	6.9	5.6	1.6	0.4	0.1	0.1
0.86	53.0	24.9	11.7	5.5	3.8	0.9	0.2	-	-
0.84	58.0	24.3	10.2	4.2	2.6	0.4	0.1	-	-
0.80	67.2	22.0	7.3	2.3	1.1	0.1	-	-	-
0.75	76.3	18.1	4.3	1.0	0.3	-	-	-	-
0.70	83.2	14.0	2.3	0.4	0.1	-	-	-	-
0.65	88.4	10.3	1.2	0.1	-	-	-	-	-

70 percent of solar radiation is permeated. Since ice, even in thin sheets, is completely non-transparent to long-wave rays, the formation of an ice crust on the surface of the snow or the soil, in the presence of solar radiation, creates conditions favorable to the thawing out of the snow and soil and the warming of the latter.

Such ice crust formations in winter-crop fields, particularly when lying directly on the grass cover and not covered up with snow, induce a premature awakening of active life in the winter crops, which may lead to their being damaged.

A temperature inversion usually prevails over the snow cover, and it becomes more pronounced in calm and clear weather (the temperature of the air over a snow cover is discussed briefly in Chapter 5). With the air temperature above zero Centigrade, which is observed over the snow cover, in the presence of warm air advection, the temperature inversion is the greater, the higher the temperature of the air. The variation in the temperature inversion in the layer between 20 and 150 centimeters in relation to the temperature at the height of 150 centimeters is represented in Figure 48. This temperature inversion is called the snow inversion, since it is tied in directly with the presence of a snow cover, the temperature of which cannot be above zero Centigrade.

Figure 48. The effect of the air temperature t_{150} upon the value of the inversion Δt_{20-150} . Figures at dots - number of observed cases.

The snow cover plays the part of a heat insulator, with the leading part played, in addition to its depth, by its density. The density of snow is the ratio of the volume of water, derived from a given volume of snow, to the said volume of snow. The density of freshly precipitated snow depends on the temperature of the air during the precipitation (Table 64, as per V. N. Obolenskiy). The higher the temperature, the higher the density of the snow.

[see Table 64 on following page]

According to B. P. Vaynberg, the density of the snow cover

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[see Table 64 on following page]

According to B. P. Vaynberg, the density of the snow cover

TABLE 64

Density of freshly precipitated snow in relation to
temperature

Air temperature during precipitation	Density of fresh snow		
	Mean	Maximum	Minimum
Below minus 10°	0.07	0.23	0.01
From minus 10 to minus 5°	0.09	0.30	0.01
From minus 5 to 0°	0.11	0.45	0.04
From 0 to plus 2°	0.18	0.53	0.07
Above plus 2°	0.20	0.59	0.16

is increased over the period of a month, in the mean by 10 percent. Table 65 cites the density of snow as of the 15th of each month, in accordance with observations by B. P. Veynberg at several locations in the USSR.

TABLE 65

Mean density of snow cover

Date	Syktynkar	Borovo forest reservation	Saratov
15 December	0.20	0.18	0.23
15 January	0.22	0.20	0.25
15 February	0.25	0.22	0.28
15 March	0.26	0.25	0.30
15 April	0.37	-	-

According to Abels, the coefficient of heat conduction of snow λ is in direct ratio to the square of its density:

$$\lambda = 0.0067 d^2 \text{ cal/cm}^2 \text{ sec degree (20)}$$

Thus, with the density of snow at 0.20, its heat conduction will be 0.0003 cal/cm² sec degree; with a density of 0.30, the heat conduction will be 0.0006 cal/cm² sec degree. For purposes of comparison, let us recall that the heat conduction of mineralized soil is 0.002-0.005 cal/cm² sec degree, i.e. 10 times higher than the heat conduction of snow, and only completely dry friable soils approximate snow, by their heat conduction value.

The covering up of the soil with snow, by virtue of which the active surface is transferred to a greater height, results in the weakening of the temperature oscillations in the soil. The low heat conduction of the snow accelerates this effect.

The monthly temperatures of snow covered and denuded soil, typical for the western part of the forest zone (as per G. A. Lynboslavskiy), are given in Table 66.

[see Table 66 on following page]

The presence of the snow cover, as can be seen from Table 66, raised the temperature of the top layers of the soil during the months of January and February ^{by} more than ~~4~~ 6 degrees. Under the snow cover, the freezing of the soil hardly reaches the depth of 0.4 meters, while denuded soil freezes to a depth in excess of 0.8 meters. In the spring, the snow cover retards

TABLE 66

Temperatures of soil under snow (I) and
denuded (II) (Mean value for 15-year period)

Months							Depth of snow
	0.0	0.1	0.2	0.4	0.8	1.6	cover in centi- meters
November I	0.1	1.3	1.9	2.8	4.5	6.0	8.0
November II	-1.6	-0.6	0.6	2.0	3.8	6.0	
△	1.7	1.9	1.3	0.8	0.7	0.0	
December I	-1.8	-0.2	0.4	1.3	2.9	4.4	21.0
II	-6.5	-5.6	-3.9	-1.9	0.9	3.9	
△	4.7	5.4	4.3	3.2	2.0	0.5	
January I	1.5	-0.4	0.0	0.8	2.1	3.4	37.4
II	-8.0	-7.3	-6.1	-4.2	-1.2	2.2	
△	6.5	6.9	6.1	5.0	3.3	1.2	
February I	-1.6	-1.0	-0.6	0.2	1.6	3.0	55.3
II	-8.8	-8.0	-6.9	-5.2	-2.3	1.3	
△	7.2	7.0	6.3	5.4	3.9	1.7	
March I	-1.2	-0.9	-0.6	0.4	1.2	2.6	60.7
II	-4.7	-4.5	-3.7	-2.9	-1.6	0.9	
△	3.5	3.6	3.1	3.0	2.8	1.7	
April I	1.8	0.9	0.8	0.9	1.4	2.3	36.3
II	3.8	3.0	2.2	1.0	0.2	0.9	
△	-2.0	-2.1	-1.4	0.1	1.2	1.4	

the warming trough of the soil, and in April the upper layers of the denuded soil turn out to be the warmer ones.

Table 67 characterizes the relation of the depth of freezing of the soil, during the month of December for individual years, to the mean monthly temperature of the air and the mean depth of the snow cover for the western part of the forest zone (as per I. A. Gol'tsberg).

It must be pointed out, that attempts to tie in the freezing through of the soil with the air temperature and the depth of the snow cover during winter months, other than December, produced indefinite results. This may be explained by the influence of other factors, including the moisture in the soil and its migration, both, in a liquid and vapor state, and also by the various densities of the snow.

[see Table 67 on following page]

Investigations of the thermal cycle of the snow cover itself were conducted by A. P. Tol'skiy at the glade station of the Borovo forest reservation Figure 49, depicting the temperature isopleths for the snow cover and the soil, by their mean values over 5-day observation periods, is a good illustration of the decrease in the temperature oscillations of the soil with the increase in the snow cover.

Soil snow cover Figure 49. Thermaisopleths in the snow cover and in the soil. Borovo Experimental Forest Reservation, winter 1908-1909 (as per A. P. Tol'skiy).

TABLE 67

Depth of the frost zone in the soil (in centimeters) during the month of December in relation to the mean monthly temperature of the air and the mean depth of the snow cover.

Mean Depth of snow cover in centimeters	Mean monthly temperature of the air											
	-1	-2	-3	-4	-5	-6	-7	-8	-9	-10	-11	-12
10	10	10	15	15	20	25	35	45	55	-	-	-
10-20	does not freeze	<5	<5	5	10	15	20	30	40	55	-	-
20	does not freeze				-	<5	<5	5	10	15	20	30

In the upper part of the snow cover, the temperature oscillations are rather considerable. A. P. Tol'skiy describes the diurnal temperature cycle in the snow cover. Figure 50 shows the isopleths of the diurnal temperature cycle in the snow cover (67 centimeters deep) as per observations in the Borovo Forest Reservation in clear frosty weather. The density of the upper 10-centimeter snow layer was 0.18, the density of the remaining layer - 0.25-0.33. The temperatures were read with the aid of specially fabricated alcohol thermometers. At the depth of 40 centimeters from the surface of the snow, the diurnal temperature cycle is leveled off. Some change in the disposition of the isopleth (-10° Centigrade), synchronized with the temperature variation on the surface, is obviously tied in with the error in the reading of the thermometers, the protruding parts of which could have been warmed by the sun.



Figure 50. Thermoisopleths in the snow cover and in the soil by hourly observations 24-26 February 1910 (by A. P. Tol'skiy).

The diurnal temperature range on the surface of the snow is 30 degrees Centigrade (36.8 degrees - 7.0 degrees). At the 5-centimeter depth of the snow, the temperature range is reduced to 16 degrees, at the 9-centimeter depth - to 11.3 degrees, at the 24-centimeter depth - to 2.7 degrees.

With the aid of this data and formula (7), the coefficient of temperature conduction of the snow cover can be determined. For the upper layer, having a density of 0.18, the coefficient of temperature conduction is $0.002 \text{ cm}^2 \text{ sec}$, and for the 9-24-centimeter layer, alongside of the increase in density to 0.26 temperature conduction also increases to $0.004 \text{ cm}^2 \text{ sec}$. This last figure coincides with the temperature conduction of soils of medium dampness. It must be remembered that, in this case, we are dealing with effective temperature conduction, combining molecular and radiation heat transfer, which is due to the pene-

tration of solar radiation into the depth of the snow cover. The evaluation of the spring warming and melting of the snow cover can be made when its effective temperature conduction is known. Waves of cold are disseminated into the depth of the snow through molecular temperature conduction only, the value of which must be below the above cited values.

The sharp increase in the temperature conduction of the snow cover with an increase in its density merits attention. This phenomenon may be one of the reasons for the difficulties encountered in the comparative study of the temperature differentials of the air and the soil under a snow cover, in relation to the depth of the snow cover. Taking into account the additional factor of the snow density will facilitate the solution of the problem.

The effect of the snow cover upon the temperature of the soil at the depth of 3 centimeters is particularly important to agricultural production. This specific depth corresponds to the node of fasciculation of the winter grain crops. The dropping of the temperature below a certain level specifically at this critical depth may be fatal to the winter crops.

A. M. Shul'gin furnishes a summary of winter temperatures of the soil at the 3-centimeter depth for the Ukrainian SSR. This data shows that a snow cover, 5-6 centimeters deep, is adequate to sustain, during a continuous period of frost stretching into a number of days, a 10-degree differential between the air temperature and the temperature of the soil at the critical depth of 3 centimeters.

The heat-insulating characteristics of the snow cover are of tremendous practical importance, since they exert an effect not only on the hibernation of the winter crops, but also on the character of the spring run-off. By preventing the soil from freezing, the snow cover at the same time promotes the percolation of the thaw waters into the soil, thereby diminishing the detrimental surface run-off. Therefore, the artificial increase in the snow cover by snow detention measures is a mighty weapon in the melioration of the climate of the soil (See Chapter 17).

DIVISION IV.

LOCAL CLIMATE

In contradistinction of microclimate, local climate, as indicated before, is determined by factors of a greater scale: mesorelief, solid masses of forest vegetation, proximity to bodies of water, construction and development. The characteristics of local climate are well discernible at the height of 2 meters and above, and can, therefore, be studied with the aid of devices and methods employed by the meteorological station network.

The local climate must be taken into account in the planning of population centers, particularly, health resorts, sanatoria, in the selection of parcels for agricultural crops in areas that are close to the climatic boundaries of said crops. Finally, our interest in local climate is linked to the possibilities existing for its modification.

We shall analyze the features of local climate by its individual control elements, with the exception of the climate of the forest and the city, which must be looked upon as local climate complexes.

CHAPTER 12

THE EFFECT OF LOCATION CHARACTERISTICS UPON THE WIND

Under conditions of interlaced relief, the air current is subject to deformation. Deformation is particularly pronounced at the terrestrial surface and ~~is~~ ^{is} fading out with ~~the~~ height, which phenomenon disturbs the normal variation of wind with height. It is manifested particularly in the disturbance of the logarithmic regularity of the variation of wind with height, when the atmosphere ^{is} in a state of equilibrium.

Figure 51 is a graph plotted in semi-logarithmic coordinates, representing the variation with height of a moderate wind (A), strong wind (B) over the plain (R), and over the hill (K1). In the first case, as was to be expected, we have a straight line, in the second case - a curve, which indicates the absence of logarithmic regularity.

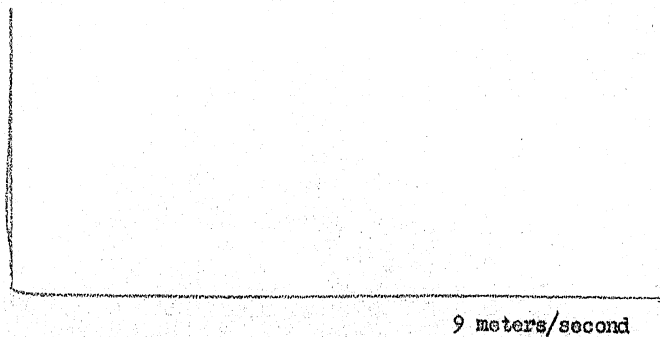


Figure 51. Deformation of the vertical profile of the wind over a hill (as per A. I. Baranov K1-hill; R - plain; A-mild wind; B - strong wind.

Let us analyze separately the effect of the relief proper and the effect of the shielding of a locality by vegetation and structures.

There are two types of air current deformation: dynamic and thermodynamic.

In the first case, we are dealing with deformation in the narrow sense of the word. An air current, encountering obstacles in its path, flows around them, there is convergence and divergence of flow lines, and, in addition, the formation of vorticity areas. With this, the velocity of the wind, i.e., the velocity of a directed air current, which is parallel to the terrestrial surface, due to the narrowing of the transverse section of the current, will increase in spots, while decreasing in other spots (Aerodynamic umbra, kinetic energy loss to friction and vorticity). Vorticities having a horizontal as well as a vertical axis are a distinct characteristic of the wind under conditions of interlaced relief.

The analysis of the structure of an air current over a locality with an interlaced relief presents great difficulties, since few direct measurements of the vertical component of the wind are available.

An indirect characteristic of the wind by-passing obstacles may be had by the settling of the snow, since the latter is easily carried by the wind.

Figure 47 in the preceding text depicts diagrams of snow accumulation and the possible structures of air currents over some elementary forms of terrestrial relief, on which the above manner of snow accumulation is dependent.

The spout of deflation, observed at the base of the windward escarpment, bears witness to the vorticity of the air current as presented in the diagram in Figure 47, A. The beak-shaped snow visor at the upper edge of the windswept escarpment, the snowdrift at its base, separated by a spout of deflation from the second snowdrift, swept up by the wind from the escarpment, also indicates the presence of a horizontal axis vortex (see Figure 47, B). Similar vorticities are also generated in a ditch with relatively shallow slope walls and on the steep slopes of hills and dunes, stretched out perpendicular to the wind (see Figure 47, V, G, D).

The formation of vortices at steep windward and leeward slopes is confirmed by instrument observations conducted by B. A. Ayzenstadt on the vertical component of the wind, it being the case that the maximum value of the rising currents over a windward slope is adapted to the boundary between the slope and the summit. Over the leeward slopes the formation of an aerodynamic umbra occurs characterized by insignificant wind velocities, which depend but little on the wind velocities over the windward slope.

To clarify the nature of this aerodynamic umbra, B. A. Ayzenstadt computed the coefficients of correlation r between the wind velocities of the windward and leeward slopes of elevations for 3 levels: 1.5, 4, and 8 meters. The results were as follows:

$r_{1.5}$ 0.45 0.11; r_4 0.59 0.09; r_8 0.78 0.06

The low values of correlation coefficients at the first two levels indicate the predominance there of a nonregulated turbulent cycle of winds. The higher value of the correlation

coefficient at the 8-meter height indicates that the zone of accelerated vorticity lies below this height.

Shallow streamlined hills with an approximately 6-degree angle of slope do not contribute to any noticeable vorticity, which is, in particular, confirmed by the higher correlation coefficients between the wind velocities of the windward and leeward slopes, having an approximate value of 0.8 - 0.9 (this does not interfere with the relative diminution in velocities over the leeward slope).

Intensive formation of vorticity most frequently occurs at the leeward slope. These vortices, at a certain phase, are usually torn away and carried off by the air current, with new vortices generated in their place. But in some cases there is the formation of steady vortices, which maintain themselves about the elevation for some longer time.

According to B. A. Ayzenshtadt, the development of steady vorticity over the leeward slope is assisted by the presence of superadiabatic gradients over the windward slope, while lower temperatures and, particularly, temperature inversions, prevail over the leeward slope.

This last phenomenon is on the boundary line between dynamic and thermodynamic deformation of the wind current.

The filtration of the air current through the obstacles, blown through by it (such as ligneous vegetation, groups of structures), a phenomenon, as yet, insufficiently studied, is to be relegated to the classification known as dynamic deformation. In these cases the air current is broken into smaller streams which

in individual spots may even manifest higher velocities, yet the total directed velocity of the air current is considerably reduced, due not only to the formation of vorticity, but also to kinetic energy losses to friction. The amount of kinetic energy lost to friction may be judged indirectly by the swaying of the trees and the damage sustained by structures.

The deformation of the air current, in the presence of interlaced terrestrial relief, runs in relation to the thermal stratification of the lowest air layers.

In the presence of superadiabatic gradients vorticity is generated even over relatively non-steep slopes. With a temperature inversion, particularly in the winter, when, due to the presence of the snow cover, the active surface is smoothened, and dynamic turbulence is abated, there is no vorticity formation even over slopes 15 or 20-degrees steep. The air current bypasses the obstacles laminaarily, with no change in its vertical stratification, it being the case that isolated elevations are bypassed predominantly along a horizontal line (from around the sides and not over the top).

In the presence of a temperature inversion, winter or summer, the lower sections of the elevation may find themselves in a zone of complete calm (provided there is no generation there of an independent circulation). Such a phenomenon was observed more than once in the Crimean State Game Reservation, at the headwaters of the

Al'ma River. The same phenomenon was observed by K. P. Kurakaya-Pakhnevich in the form of nocturnal interruptions of the dry winds (a sudden calm, an abrupt temperature drop, and a rise in relative humidity) in the deep valleys between the foothills of Western

Transcaucasia, with the round-the-clock dry wind continuing in full force over the adjacent exposed sections. The lighter warm and dry air is sliding over the valley and is prevented from penetrating into it by the heavy cold air, with which it is filled.

Such air stagnations may also be observed over glades in a forest. The height, to which the motionless air reaches, is determined by the height of the surrounding elevations and vegetation, but also runs in relation to weather conditions and to the height of the temperature inversion layer. Wind acceleration will decrease the height of the stagnant air, sometimes even destroy it completely, while wind deceleration leads to an increase in its volume.

We now pass to quantitative analysis of wind velocity, in the presence of interlaced terrestrial relief.

The quantitative ratios between the wind velocities at the height of 2 meters, over various forms of terrestrial relief, that are typical of the topography of a plain, were obtained by observations of the Central Geophysical Observatory expedition in the Saratov area. They are presented, in their averaged form, in Table 68.

See next page for Table 68

TABLE 68

Wind velocity at the height of 2 and 10 meters, in relation to the form of terrestrial relief, expressed in ratios to the velocity of the wind over an exposed level¹ place.

w $\frac{u_0}{v_0}$, with the wind velocity over the level ground over 2 meters/second

Form of Relief		w_2	w_{10}	Characteristic of Slopes ²
Top of a steel elevation		1.4 - 1.6	1.3 - 1.4	Steepness of 7 - 12°
Top of a shallow elevation		1.1 - 1.2	1.1	Steepness of 4 - 6°. Difference in heights 10 - 50 meters
Windward and parallel to the wind slopes of hill	top middle bottom	1.3 - 1.5 1.1 - 1.2 1.0	1.2 - 1.4 1.1 1.0	Upper part of slopes 5 - 10 meters below the top. Accelerated dip of slope 6 - 12°
Leeward slopes of hill	top middle bottom	1.2 0.9 0.6	1.1 0.9 0.7	Middle part of slopes below top by 10 - 20 meters Uniform dip of the slopes 8 - 10° Lower part of slopes
Depression		0.9	0.9	Decelerated dip of slopes 3 - 6°
Gully		0.5 - 0.6	0.6 - 0.7	Slopes of gully 3 - 4° steep

¹A level location is a parcel of land with an angle of inclination less than 2 degrees, occupying an area with a radius of 100 meters, with a distance to the nearest relief irregularity (hill or escarpment) equal to at least 20 times its elevation.

Footnotes to Table 68 (Continued)

²The quantitative analysis of the terrestrial relief is given on the basis of the relief characteristics at the observation point, and its purpose is reduced basically to a more precise specification for the limits of possible interpolation of the wind characteristics.

The wind velocities in the air layer from 1 to 5 meters high, over a hill and over a gully, expressed in ratios to the wind velocity over an exposed level parcel of land, is presented graphically in Figure 52.

Wind Velocities
Higher than
Over Plain

Direction of Wind

Wind Velocities
Higher than
Over Plain

Figure 52. The variation in the wind velocity in the 1 - 5 meter layer, expressed in ratios to the corresponding nocturnal velocities over level ground, with the velocity at the height of 2 meters being 2 - 4 meters/per second.

A - hill; B - depression with gully

The elevation of the hill - 20 - 30 meters; slope steepness - 6 - 10 degrees; depth of the depression 20 - 30 meters; steepness of its slopes 6 degrees; depth of the gully - 5 meters.

Data collected by the same expedition made it possible to establish that, in its first approximation, the effect of the form of terrestrial relief is a logarithmic function of the height, i.e.

$$\frac{w_1 - w_2}{w_2 - w_3} = \frac{\lg z_1 - \lg z_2}{\lg z_2 - \lg z_3}, \quad (21)$$

where w_1 , w_2 , and w_3 are the ratios of wind velocity under the given conditions of relief at heights z_1 , z_2 , and z_3 to the velocity of the wind over level ground at the same height. In other words, the effect of the terrestrial relief is diminished in direct ratio to the logarithm of height.

If the ratio of wind velocity, under given conditions of relief, at the height of 2 meters, to the wind velocity over exposed level ground, is known, then the corresponding velocity ratios at the heights of 1, 5 and 10 meters (w_1 , w_5 , w_{10}) can be derived by the following formulas:

$$w_1 = 1.2w_2 - 0.2, \quad (22)$$

$$w_5 = 0.8w_2 + 0.2, \quad (23)$$

$$w_{10} = 0.7w_2 + 0.3. \quad (24)$$

The wind velocity ratios at the height of 10 meters, cited in Table 68, were derived by this formula.

The quantitative analysis of the effect of the location characteristics upon wind velocity could also be obtained by the data of the meteorological stations. The first such attempt was made by M. O. Podtyagin in 1955. But Podtyagin in his "classes" was making a summary evaluation of a location, without separating

the effect of individual features, such as relief, vegetation, structures, and other factors. His criterion for relegating a location to a certain class was the wind velocity itself.

The basic defect of such a summary evaluation is the difficulty in arriving at a physical interpretation of the contemplated phenomenon.

According to the data of the meteorological stations, a great effect upon wind velocity is exerted, in addition to relief, by the degree to which locations at various heights are sheltered by trees and buildings.

By the degree of shelter, all meteorological stations can be divided into two groups: (1) with the sheltering obstacles below the height of the anemoscope; (2) with the sheltering obstacles above the height of the anemoscope.

In relation to the number of sheltering obstacles, these two groups are divided into types.

The type classification of a location by the degree of shelter is depicted in Table 69.

[See next page for Table 69]

TABLE 69

Location Characteristics of Stations by Degree of Shelter

Anemoscope above sheltering objects	1. Exposed location, individual structures and trees
	2. In the midst of a village, garden, forest plantings
Anemoscope below sheltering objects	3. Trees and structures above anemoscope
	4. Anemoscope sheltered all around by structures and trees

Note: Sheltering objects are considered as such when their distance from the anemoscope is not less than 20 times their height.

The mean annual wind velocities, by mean diurnal values, as well as monthly schedule annual temperature ranges (ratio of maximum mean monthly velocity to the minimum mean), and maximum diurnal ranges (ratio of velocity at 1300 hours to velocity at 2100 hours), for the first 3 types of stations in the middle latitudes of the European part of the territory of the USSR, are cited in Table 70. The annual and diurnal ranges are given not in differentials, but in ratios, proceeding from the fact that, in theoretical and practical computations, wind velocity is usually taken as a factor and not as an item, and, consequently, it is important to know not its absolute, but its relative variations.

TABLE 70

The effect of the sheltering of a location upon wind velocity (meters/per second) over level ground (converted to the height of 10 meters)

Degree of Sheltering of a Location	Mean Annual Velocity				Annual Range				Maximum	Number of Stations
	Mean	0700	1300	2100	Mean	0700	1300	2100	Diurnal	
	Diurnal	Hours	Hours	Hours	Diurnal	Hours	Hours	Hours	Range	
1. Exposed location: individual trees and structures (u_1)	4.4	4.2	5.3	3.7	1.5	1.6	1.3	2.1	$\frac{u_3}{u_2}$	9
2. In the midst of a village, garden, forest plantings below 10 meters (u_2)	3.3	3.0	4.2	2.8	1.5	1.6	1.5	2.1	2.0	8
$\frac{u_2}{u_1}$	0.75	0.71	0.79	0.76						
3. In a village, garden, forest plantings, individual trees and structures above 10 meters (u_3)	2.9	2.6	3.5	2.6	1.5	1.7	1.3	2.0	1.8	4
$\frac{u_3}{u_1}$	0.66	0.62	0.66	0.70						

It follows from Table 70 that the mean wind velocity at the stations of type 2 amounts to 75 percent of the velocity over an exposed location. In the case of the third type of station, it is 66 percent.

Particular attention should be paid to the uniformity of the annual and diurnal wind velocity variation regardless of the station type. This furnishes a basis for assuming that the degree of stability of the atmosphere plays no substantial part in the effect exerted by the sheltering characteristics of a locality, since the atmospheric stability differentials specifically form the diurnal variation, but at the same time cause the deformation in the annual variation of wind velocity.

The data of meteorological stations located in various conditions of terrestrial relief, in the presence of a uniform degree of sheltering, is cited in Table 71.

[See next page for Table 71]

In synthesizing Tables 68, 70, and 71, we derive a summary table (Table 72) which characterizes the variations in the mean annual wind velocity in relation to terrestrial relief and the degree of sheltering of a locality, with the addition of the characteristic of the diurnal wind velocity variation. Naturally, Table 72, compiled by data pertaining to the central part of the European territory of the USSR, requires additional checking when applied to other territories.

TABLE 71

The effect of terrestrial relief upon wind velocity (meters/second), as per observations of stations located in villages, gardens, and forest plantings, in ratios to the wind velocity over a plain (u) (converted to a 10-meter height)

Form of Relief	Mean Annual Velocity				Annual Range				Maximum	Number of Stations
	Mean	0700	1300	2100	Mean	0700	1300	2100	Diurnal	
	Diurnal	Hours	Hours	Hours	Diurnal	Hours	Hours	Hours	Range	
									$\frac{u_{13}}{u_{21}}$	
Shallow summit u_v	4.0	3.7	4.8	3.6	1.4	1.4	1.3	1.9	1.8	4
$u_v : u_e$	1.14	1.15	1.12	1.20						
Shallow slope valley u_d	3.2	3.0	4.2	2.5	1.6	1.9	1.6	2.9	3.1	3
$u_d : u_e$	0.92	0.94	0.98	0.76						
Deep valley u_{gd}	2.2	2.1	2.7	1.7	1.3	1.3	1.6	2.4	3.1	1
$u_{gd} : u_e$	0.63	0.66	0.63	0.57						

TABLE 72

Wind velocity (in meters/per second) in relation to relief and degree of sheltering of a locality (at the height of 10 meters)

Mean Annual Wind Velocity

				Anemoscope	
				Below Sheltering Objects	Maximum differential range
Anemoscope Above Sheltering Objects					
				Objects	$\frac{u_{13}}{u_{21}}$
Form of Relief	exposed place,	in village,	in village,		
	individual structures and trees	garden, forest plantings	garden, forest plantings		(does not depend on degree of sheltering)
Negative	} steep	3.3	2.5	2.2	3.1
	} shallow	4.0	3.0	2.6	3.1
Level Place		4.4	3.3	2.9	2.1
Positive	} shallow	4.8	3.6	3.2	1.8
	} steep	5.5	4.1	3.6	--

It follows from all the above-said that in utilizing wind velocity data from climatological manuals it is necessary to be guided not by the proximity of the meteorological station to the contemplated location, but by the uniformity of conditions at the locality.

It is now in place to add a few words about the variation

in the direction of the wind under conditions of interlaced relief. A change in the direction of the wind by one rhumb (22.5 degrees) is the usual phenomenon under interlaced conditions of relief, it being the case that in the winter, under the same conditions of relief, the departures are greater (up to 2 rhumbs -- 45 degrees). As a result, over a valley, the number of winds, blowing along the valley, will increase, while near structures, masses of vegetation, and individual elevations, the recurrence of winds from the side of the sheltering objects will be decreased, it being the case that the recurrence of winds of adjacent rhumbs is usually increased.

As an example, the wind-roses of two meteorological stations located side by side (Nos 5 and 6) in the Velikoanadol'skoye Forest Reservation are shown in Figure 53. One of the stations is located at the northern edge of a forest, the second one in the open steppe. While over the open steppe, the prevailing wind was from the north, a southerly wind was observed over the northern edge of the forest.

Figure 53. Wind-roses. Velikoanadol'skoye Forest Reservation, March 1893. I - northern edge of forest; II - open steppe.

We will now pass on to the analysis of independent local air currents of a thermodynamic origin. These winds, stipulated by the differences in the temperatures of the air over the slope and the valley, over the land and the body of water, over the forest and the field, at times completely overlap the basic wind, which is determinable by the general circulation of the atmosphere.

Local winds in areas of interlaced relief, with prevailing differences in altitude less than 200 meters, which is typical of the European territory of the USSR, have, as yet, been studied very little. Some conception of the local winds may be formed by the juxtaposition of the wind-roses of meteorological stations located in direct proximity to each other. As an example, the wind data of 3 stations located at the elevated plain near the Volga River is shown in Figure 54. During the day, the wind-roses of the 3 stations are similar, while at night there is a sharp divergence.

1500 hours

1400 hours

Scale 10 percent

Figure 54. Wind-roses, June-August. A - Shikhary (1943);
B - Privol'skaya (1941 - 1942); V - Sinodskoye (1941 - 1942).
The number of calm occurrences is given in percent.

The absence of daylight local winds, noticeable under conditions of small altitude differences, is not just casual. The difference in the characteristics of the daylight and nocturnal

air current is linked directly with the corresponding difference in turbulence.

The considerable turbulent exchange during the day rapidly levels off the small (by comparison with a mountainous area) thermal differentials over a locality with interlaced relief, and, by virtue of this, liquidates the sources of the local winds. The abated nocturnal turbulent exchange lacks the strength required for this leveling off. This fact is confirmed by the commonly known greater variability, under the effect of local conditions, of the nocturnal temperatures, as compared to daylight temperatures.

At night, the air, that has been cooled at the surface and is, therefore, heavier, flows down. Local nocturnal winds are characteristic for valleys, hollows, elevation slopes in all cases where the general angle of slope of the locality is more than 3 degrees. The nocturnal wind is originated not at the upper part of the slope, but at some distance from it, since, in order for a local air current to come into existence, the presence of a corresponding air accumulation area is a prerequisite. This means relatively large in extent and altitude differentials forms of terrestrial relief. The height of the local wind is a certain, as yet undetermined, function of the relief altitude differentials existing in a locality, and it may oscillate within considerable limits. In the presence of a small difference in relief altitudes, the power of the air current may be very insignificant. Thus, at the slope of the same elevated plain near the Volga, at a point 40 meters below the summit, the shift from the local air current to the basic wind is noticeable already at the

height of 15 - 20 meters.

In the intermediate layer, a sort of wind inversion may be observed, i.e., a decrease in wind velocity with height.

A condition prerequisite to the formation of a local nocturnal wind is calm and clear weather, which provides for the radiation chilling of the air.

In mountainous areas, not only nocturnal mountain breezes, but also daylightvalley breezes are observed (Below, there is only general data on mountain-valley breezes, since detailed characteristics are given in general courses on meteorology and climatology.)

Mountain-valley circulation, as per A. Kh. Khrgian, is composed of slope winds and the "mountain breeze", which are stipulated by the difference in the thermal balance prevailing in the upper and lower parts of the valley.

A substantial moment in the evolution of the slope wind during the day is the greater warming of the air over the slopes as compared to the free atmosphere at the same height. Thus, a reservoir of warm air is created along the slopes, while, at the same time, there is a source of cold air in the free atmosphere. As a result of this, the isobaric surfaces in the direction of the slopes are lifted, creating aloft a pressure gradient, directed toward the valley.

In connection with the evolving flow-off of the air from the summit to the valley, a pressure gradient in the opposite

direction is created below, stipulating the wind along the slope and upward. In their particularly strong development during the meridional hours, the slope winds induce the formation of cumuliiform clouds (see Chapter 14). The daylight slope winds are related to the warming of the slope, and, therefore, attain higher velocities over southern slopes as compared to northern. In the upper part of the southern slopes in one of the valleys of the Gissarek Mountain Range (in the Botanical Garden of the Tadzhik Affiliate Academy of Sciences), the velocity of the wind during the summer daylight hours was 6 meters/second, while over the northern slopes it was only 2 meters/second.

During the night, the air at the slope is colder than the air in the free atmosphere, leading to a reverse type of circulation, i.e., at the terrestrial surface, the wind blows down the slope. It being the case that over the lower part of the slope the wind carries the air, which was cooled considerably in its movement along the surface, with high relative humidity, while, to the upper part of the slopes, the air flows from the free atmosphere and, due to the descending motion, it frequently has a relatively high temperature and a low relative humidity. This shift in the daylight and nocturnal wind is the determining factor in the peculiarity of the diurnal variation of relative humidity in the mountains, with the minimum value occurring not during the day, but at night.

Under "mountain breezes", or mountain-valley winds, in the narrow sense of the word, are understood winds developing in relatively wide and slowly rising valleys. This phenomenon, as yet, has been studied little. It is stipulated by the difference in the

heating and the cooling between the upper and lower part of the valley. Mountain-valley winds are linked directly to the slope winds. During the day, the valley wind (up the valley) compensates for the flowoff of the air up the slope, which is accomplished by the slope wind. The nocturnal mountain breeze carries down the valley the cold air descending from the slopes in the form of the nocturnal slope wind.

Observations by A. Kh. Khragian in the Tseyskoye Canyon of the Northern Caucasus resulted in the following analysis of the mountain-valley breezes.

In the morning hours, the slope winds are more pronounced. During the day they take second place, the predominance shifting to the valley wind. The vertical depth of the valley breeze is about one kilometer. Over the flow of the valley breeze, there is always present an intensive reverse current.

The mountain breeze, becoming stronger during the second half of the night and continuing after sunrise, is milder and less stable than the valley breeze. Its vertical depth is also considerably below that of the mountain breeze. The reverse current over the mountain breeze is sometimes stronger than the breeze itself.

The reciprocal action between the slope winds and the mountain-valley breezes frequently causes considerable variations in the diurnal wind cycle. Over the center of narrow valleys, the direction of the wind shifts abruptly by 180 degrees. At the foot of wide valley slopes, the ascending slope wind appears 1 - 2

hours ahead of the valley breeze, and is then completely displaced by it. In the evening, taking over from the latter, the descending slope wind appears on the scene to be later, in turn, displaced by the mountain breeze. As a result, there is a clockwise rotation of the wind at the left-hand valley slopes (looking down the valley), and an anti-clockwise rotation at the right-hand valley slopes.

A few words about the glacier breeze -- which is the flowoff of cold air along a glacier without a change in direction, day or night, inasmuch as the temperature of the glacier surface is always below that of the air. Its depth is usually not great, from several tens to 100 meters.

According to the expression used by A. Kh. Khragian, who observed the wind cycle of the Tseykiy glacier, the glacier and the valley breezes are antagonists.

The glacier wind forms a cold wedge, which, in creeping down from the glacier, displaces the valley breeze upwards. At the lower boundary of the glacier wind, there is frequent formation of fogs, which, in the opinion of A. Kh. Khragian, are of the mixed fog variety, since, in this case, there is indeed an energetic intermixing of the warm valley and cold glacier air. The valley breeze somewhat retards the evolution of the glacier breeze in daylight, therefore, the peak value of the glacier breeze is adapted to the nocturnal hours, when its direction coincides with the direction of the mountain breeze.

At the boundary of bodies of water, occurs the evolution of a breeze circulation, embracing an atmospheric layer of a certain

depth. During the day, when the land mass is warmer than the sea, a sea breeze is evolved, which is made up of a lower breeze current, directed from sea to land, and an upper current, directed from land to sea, the current being ascendant over the land, at the breeze boundary, and descendant over the sea. The reverse is true in the case of a nocturnal breeze occurring over a littoral: the lower breeze current is from land to sea, while the upper one -- from sea to land.

The theory and the detailed characteristics of the breeze circulation are to be found in general courses on meteorology and climatology. We will limit ourselves here to the analysis of the effect of local conditions upon the evolution of breezes.

The greater the difference in the temperature between the surfaces of the water and the land mass, the more frequent and intensive is the evolution of breezes. Therefore, over littorals covered either with swamps or abundant vegetation, which loses great quantities of heat to evaporation, the daylight breeze will be abated. On the contrary, over desertlike rocky littorals, the daylight breeze is distinct by greater intensity.

Both the daylight and the nocturnal littoral breeze is intensified in those cases, when it becomes a valley-mountain breeze, i.e., at the places of egress into the littoral of mountain valleys.

The effect exerted by terrestrial relief is well illustrated by the data on the recurrence of breezes in the Crimea, assembled by V. Kellerman (Table 73).

TABLE 73

The mean annual recurrence of breezes (number of cases)

Observation Point	Day	Night
Saki	116	122
Sevastopol'	148	115
Sarych	125	121
Yalta	194	190
Sudak	190	190

As can be seen from Table 73, almost under homogeneous climatic conditions, the number of breezes varies by more than time and a half. The breezes occur most frequently in Yalta and in Sudak, where the valleys (with a difference in altitude in excess of 500 meters) emerge onto the very coast. Near Cape Sarych, the "yayla" [rock formation?] approaching the coastline in the form of a precipitous wall, somewhat hinders the evolution of breezes. Sevastopol', located in the foothills of the Crimea (with an elevation difference of 100 - 200 meters), occupies an intermediate position. And, finally, the smallest number of breezes occurs at Saki, which is due to the network of lakes, interfering with the contrast between the sea and the land.

In order to clarify the same problem, P. A. Vorontsov utilized the triple-standard aerological observations conducted simultaneously at three locations on the Black Sea. The first location was at a narrow littoral, at the foot of steeply rising rocky mountain ridges, with a southwestern exposure, running

parallel to the coast. The second location was a lowland extending several kilometers into the sea and covered with a forest, beyond which there were hills. The third location was at a low-lying swampy valley coast.

Data characterizing the depth of the breeze current in relation to nature of the littoral (as per P. A. Vorontsov) is depicted in Table 74.

TABLE 74

Depth (in kilometers) and recurrence of the breeze
current

Characteristics of the breeze current	Point 1	Point 2	Point 3
Depth of the lower current of the sea-to-land breeze	0.7	0.4	0.2
Recurrence over period of observation	16	8	2
Depth of the lower current of the land-to-sea breeze (2000 hours)	1.3	1.2	0.7
Recurrence over period of observation	13	8	1

The breezes attain their maximum depth and recurrence over the mountain slopes exposed to the southwest (Point 1). Forest overgrown lowlands occupies the intermediate point. Over a swamp covered littoral, breezes (night or day) occur very rarely (Point 3).

Breezes also in the areas of large lakes, such as Lake Ludoga, Lake Onega, Lake Sevan.

A recurring shift in the direction of the wind on the nature of a breeze occurs along the shores of large rivers. V. Kellerman indicates that in Saratov, which is located on the high right bank of the Volga River, there is the recurrence of a breeze 65

days during the year.

It must be remembered that small bodies of water are warmed through rapidly, therefore, the day breeze over them occurs at the beginning of the summer only.

Local winds, mountain-valley, as well as breezes, exert a great effect upon the formation of the local climate as a whole, including the thermal cycle, to the analysis of which the next chapter is devoted.

CHAPTER 13

THE EFFECT OF THE PECULIARITIES OF A LOCATION UPON
ITS TEMPERATURE AND HUMIDITY CYCLE

In the formation of local peculiarities of the thermal cycle and air humidity, in addition to the nature of the active surface and the radiation cycle, an important part is played by the wind, both, directly as well as indirectly through the turbulent exchange controlled by it.

As far back as 1881, A. A. Voyeykov pointed out that the diurnal temperature range in valleys is usually greater than over plains, over plains - greater than over slopes and summits, since, during the day, the mountain tops are not warmed through as much as the valleys, due to an accelerated emission of heat into the free atmosphere, and, during the night, they are not chilled as much as the valleys, since the cold air descends into the valley. In the valleys - the nights and early mornings are cool and damp, the afternoons are warm and dry. Over the slopes and the mountain tops, on the contrary, the nights are considerably warmer, the days considerably cooler than in the valleys.

Examples of the diurnal temperatures variation under various conditions of terrestrial relief in the Amurskaya Oblast are cited by P. I. Koloskov. At the Pikan Experimental Station, simultaneous observations were made in an exposed valley, on an embankment, and on^a mountain rising 343 meters above the valley. The result of these observations are presented graphically in Figure 55.

Winter

Summer

Figure 55. Diurnal temperature variations, Pikan (as per P. I. Koloskov). - valley; --- embankment; -. -. -. mountain.

During the summer, the diurnal temperature range on the mountain was only half the range in the valley, due to a higher minimum and a lower maximum. During the winter, 'round the clock, the temperature on the mountain was higher than in the valley, with the temperature on the mountain maintained on the same level, while in the valley there was a 10° variation due to the daylight disintegration of the nocturnal temperature inversion.

The peculiarities of the diurnal temperature variation are also manifested in the mean monthly temperatures of the air, as can be seen from Table 75 (as per P. I. Koloskov), in which

the temperature differentials valley-mountain, valley-embankment are cited.

TABLE 75

Mean monthly temperature differentials valley-mountain
(I) and valley-embankment (II) Pikan, 1914

	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
I	-6.1	-8.0	-2.1	1.1	1.2	0.8	-0.2	-1.1	-1.0	-1.3	-1.7	-6.4
II	3.1	-4.8	-3.0	-0.2	-0.4	-0.3	-0.5	-0.8	-0.6	-1.2	-1.7	-4.8

The effect of the terrestrial relief upon the temperature of the air is in direct relation to the velocity of the wind and the degree of cloud formation. Data characterizing the variation in the nocturnal temperature differentials valley-mountain, in the mean value for the entire period of observations, in relation to wind velocity and the degree of cloud formation, is given in Table 76 (as per P. I. Koloskov)

[see Table 76 on following page]

Due to the wind and cloud formation, the temperature differentials valley-mountain are not only diminished, but even change algebraic signs. It becomes colder on the mountain, and the temperature drop with height attains a value of 1 degree per 100 meters.

TABLE 76

Nocturnal temperature differentials valley-mountain
in relation to cloud formation and wind velocity, Pikan,
1913-1914

Degree of cloud formation designated in balls	Wind velocity in meters/second			
	0-1	2-4	5-9	10
0-2	-11.7	-4.7	1.8	
3-7	-8.0	-2.9	2.1	3.4 ¹
8-10	-1.9	1.8	2.6	

1. Regardless of cloud formation.

It should be pointed out that the mean annual wind velocity in the valley was 2.1 meters/second, i.e., considerably less than in the similarly exposed shallow valleys in the European territory of the USSR (see Table 72). This is due to the general diminution in wind velocity in the Amurs-kaya Oblast.

P. I. Koloskov also characterized the effect of terrestrial relief upon the temperature of the air by the observations of the meteorological stations of the Kazakh SSR (Table 77)/

[see Table 77 on following page]

In the European territory of the USSR, and also in Western Transcaucasia, the effect of terrestrial relief upon the thermal cycle is less pronounced. The causes of this phenomenon lie not

TABLE 77

The effect of terrestrial relief upon the temperature of
the air (in departures from temperature over a
level place

Form of relief	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
Slopes	3.9	4.1	2.5	0.3	0.1	-0.1-0.7	-0.7	-0.2	0.9	1.8	2.6	
Summits	3.0	2.6	1.5	1.3	0.9	1.3	1.7	1.7	2.2	2.4	2.7	2.0
Valleys	-3.3	-3.4	-2.0	-0.8	-0.8	-1.0	0.2	-0.3	-0.9	-1.7	-2.2	-2.2
Synclines	-9.6	-7.8	-5.2	-2.2	-3.1	-1.3	-1.7	-2.6	-2.9	-2.4	-4.2	-3.4

only in the nature of the terrestrial relief, but also in the weather conditions - higher wind velocities and a greater degree of cloud formation.

Table 78 shows the mean monthly temperature differentials of two stations in Western Transcaucasia. One of them - Makharadze, lies in a wide valley, the second one - Anaseuli, on a hill rising over the valley to a height of 90 meters (similar to the embankment in Pikan). The temperature differentials, as we can see, oscillate within the limits $\pm 0.8^\circ$.

[see table 78 on following page]

During the cold season of the year, the decisive effect upon the mean monthly temperature differentials is exerted by the

TABLE 78

Mean monthly temperature differentials between
valley (makharadze) and hill (Anaseuli)

I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
-0.7	-0.4	0.5	0.1	0.1	0.2	0.8	0.4	0.0	-0.9	-0.5	-0.8

nocturnal temperature inversions, and during the warm season - by the overheating of the valley during the day.

In the canyons sheltered by the precipitous slopes from the heat of the sun, during the day, and from the emission of heat, during the night, the diurnal temperature oscillations are considerably smaller as compared to those over the relatively more shallow valleys and synclines.

The proximity of bodies of water has a substantial effect upon the thermal cycle of the air over the adjacent land areas. The mechanics of this effect is elaborated upon in the general courses of meteorology and climatology. Here, we will limit ourselves to the quantitative analysis of the phenomena, taking into account the part played by the land mass itself, the nature of its humidification.

The direct proximity of bodies of water changes the diurnal temperature range. On the shore, as well as on the hills, daylight

temperatures are lower, and nocturnal temperatures higher as compared to areas further away from bodies of water.

I. A. Gol'tsberg (Table 79) gives quantitative evaluation of the effect exerted by the proximity of large lakes, and by the form of terrestrial relief, upon the diurnal temperature range, by the data collected by the meteorological stations in north-western forest zone of the European territory of the USSR (the + sign means that the temperature range is higher, the - sign means that it is lower than the temperature range of an exposed level locality far away from the body of water.

TABLE 79

Location	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
Littorals) islands)	-0.5	-0.5	-0.5	-3.5	-4.0	-4.0	-3.5	-3.0	-2.5	-1.5	-0.5	-0.5
Glades	+0.5	-1.0	+1.0	0.0	+0.5	+0.5	+1.0	+1.0	+1.0	+0.5	0.0	+0.5
Valleys and lowland swamps	+0.5	+0.5	+1.0	+0.5	+1.0	+1.5	+2.5	+2.5	+2.0	+1.5	+1.0	+1.0

As can be seen from Table 79, the effect of the proximity of bodies of water is considerably abated during the winter months, when the water is frozen over.

The effect of the proximity of bodies of water upon the

thermal cycle of the vegetation period in the north of the European territory of the USSR (Karelia and the Kola Peninsula is depicted in Table 80).

TABLE 80

The effect of bodies of water upon the thermal cycle of the air over littorals

Body of water	Distance from water	July temperature	Number of days with temperature above 10° C.	Date	
				Begin-ning of 5°	End of 5°
Lake Onega	Lakeshore island	16.4 15.4	105 96	4 V 15 V	6 X 10 X
White Sea	60 Kilometers Sea coast	14.6 13.0	79 66	18 V 25 V	29 IX 28 IX
Lake Imandra	1.5kilometers	13.9	71	22 V	20 IX
	0.2 kilometers	13.2	65	26 V	21 IX

Similar observations were made over the Black sea littoral in Western Transcaucasia (Table 81).

[see Table 81 on following page]

Based on detailed observations of the climate of Kazakhstan, P. I. Koloskov analyzes the effect of the bodies of water upon the

TABLE 81

The effect of the Black Sea upon the
thermal cycle of the air over the littoral

Distance from the Sea	July temperature	Number of days with temperature above 15° Centigrade	Date	
			beginning of 10°	end of 10° temperature
Sea Coast	22.9	169	26/III	25/XI
60 kilometers	23.5	184	19/III	28/XI

air temperature there (Table 82). First, a determination, by the existing maps, was made: (1) of the difference in the temperatures on the shore and at the boundary of the areal effect; (2) of the distance from the shore to the boundary of the areal effect.

[see Table 82 on following page]

The Caspian Sea exerts the greatest thermal effect, particularly in an eastern direction. The effect is warming from November to February inclusive, and cooling from March to September. The maximum warming effect, in the case of all bodies of water is timed to the month of January, with the exception of Lake Chelkar, which, by this month, is frozen over solid.

By comparing Tables 80 and 81, we can make the deduction that the effect of the proximity of a water body runs in relation to the

Table 82

The effect of some bodies of water in Kazakhstan
upon the temperature of the air.

Month	Temperature differential water-land					Width of the zone of temperature effect(in kilometers)			
	Caspian					Caspian			
	Sea area		Sea of Lake		Lake	Sea area		Sea of Lake	Lake
	over water	over land	Aral	Balkhash	Chelkar	over water	over land	Aral	Balkhash
I	5.0	4.0	2.0	2.8	0.3	275	500	75	} up to 50 up to 10
II	3.0	2.0	0	2.8	0	250	500	-	
III	-6.0	-2.0	-2.0	-1.0	-1.0	250	125	150	
IV	-4.0	-2.0	-1.5	-1.5	-1.5	300	150	100	
V	-3.0	-2.0	-1.5	-1.2	-1.8	300	200	50	
VI	-3.5	-2.0	-2.0	-0.8	-1.0	300	200	100	
VII	-4.0	-2.5	-2.0	-1.3	-1.0	300	250	75	
VIII	-2.5	-2.5	-1.5	-1.0	-0.5	300	250	50	
IX	-1.5	-1.5	0	1.8	0	200	200	-	
X	0	0	0	2.5	0	-	-	-	
XI	3.0	0	1.0	3.0	0.5	300	-	75	
XII	3.5	2.0	1.5	3.0	0.5	300	500	75	
Year	-0.8	-0.5	-0.5	0.8	-0.5	-	-	-	
Range	11.5	6.5	4.0	4.5	2.3	-	-	-	

characteristics of the land adjoining the water. For example, in the presence of the swamp-covered littoral of Western Transcaucasia, the effect of the sea is at its minimum, while on the desert-like littorals of the Caspian and Aral Seas, the effect of the water is particularly pronounced, since, under these conditions, the differentials in the thermal balance as between land and sea are particularly great.

The most substantial effects of bodies of water in the direction of reducing temperatures are timed to the spring. This is manifested even by the slowing up in the rate of growth of vegetation. If the body of water freezes over solid during the winter, its maximum warming effect takes place prior to freezing over, i.e., at the beginning of the winter. Non-freezing bodies of water exert their maximum warming effect during the month of January.

The effect exerted by a body of water fades out with distance, it being the case that, in its first approximation, the rate of fade-out may be considered as in direct ratio to the logarithm of the distance from the water.

The logarithmic character of the diminution in the cooling effect of the Caspian Sea upon the spring (April) and summer (July) temperatures is presented in Figure 56. The graph is plotted from the data collected by 13 Azerbaydzhan stations, located on the Apsheron Peninsula, in the valley of the Kura and its tributaries. The data is arranged by distance intervals (less than 0.5 kilometers, from 0.5 to 5 kilometers, from 5 to 50 kilometers, and above 50 kilometers). The disposition of the points in Figure 56 along a straight line, as plotted by

semilogarithmic coordinates (departure less than 0.2°) confirms the logarithmic character of the ratio between temperature distribution and distance from the sea. A similar graph further in the text - Figure 60) characterizes the increase in the frost hazard in relation to the logarithm of the distance from the sea.

July

April

Figure 56. Mean monthly temperature variation with distance from the sea (in semilogarithmic coordinates), Eastern Transcaucasia.

The effect of bodies of water upon the air temperature of islands and the windward slopes of surrounding elevations is particularly strong. As an example, the effect of Lake Baykal can be cited. The mean monthly temperature of the islands and the littoral are given in Table 83 (according to A. V. Voznesenskiy).

Table 83

Mean monthly temperature of the air (Lake Baykal)

Location	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
Island	-16.5	-17.6	-11.6	-2.8	4.4	10.6	12.5	13.5	9.0	1.6	-4.0	-9.7
Lake Shore	-17.2	-17.5	-11.4	-1.6	4.6	10.9	14.0	14.3	8.6	1.3	-5.3	-12.4
25 kilometers from shore	-25.8	-22.7	-12.5	-0.1	8.3	16.5	18.8	16.4	9.1	-0.6	-11.2	-24.4

Over the island, the annual temperature range is only 31.1 degrees Centigrade, while 25 kilometers away from the lake shore it is 44.6 degrees. In addition, over both stations of Lake Baykal, the maximum and minimum of the annual temperature cycle lag behind by one month. The temperature differential is particularly great during the month of December, when the lake is not yet frozen over. However, the most sensitive temperatures for vegetation are the lower (by 3 - 6 degrees) summer temperatures. As a result, a peculiar vegetation zone inversion is observed on the slopes surrounding Lake Baykal: forests in the middle part of the slopes, bordered from above and from below by a subalpine belt.

P. I. Koloskov succeeded in clearing up the effect of the degree of dampness of the soil upon the air temperature.

The greater the dampness of the active surface, the greater, all other factors being equal, is the evaporation of water, i.e. the greater the loss of heat to evaporation, which, in turn, must

result in a relatively lower air temperature. P. I. Koloskov classified the Kazakhstan meteorological stations into 4 groups by the degree of dampness prevailing in the surface of the soil, as indicated by the vegetation cover, also taking into account the availability of irrigation. He then made a comparative study of the temperatures observed and the temperatures indicated on existing maps (specific features of individual stations are disregarded in the construction of the maps). These temperature differentials are cited in Table 84.

TABLE 84

The effect of moisture in the soil on the temperature of the air (differentials between observed temperatures and temperatures indicated in existing maps)

Character of damping	V	VI	VII	VIII	IX	Mean
Dry areas	0.9	1.5	1.6	1.7	1.7	1.5
Areas with moderate natural damping	0.4	0.9	-0.3	0.1	-0.4	0.1
Areas with consider- able natural damping	-1.3	-0.7	-1.3	-1.6	-0.7	-1.1
Areas with moderate irrigation	-0.1	-0.6	-1.7	-1.3	-0.6	-0.9
Areas with inten- sive irrigation	-1.5	-1.6	-3.1	-2.6	-1.4	-2.0

By comparing the data in Tables 84 and 82, we come to the conclusion that, during the summer, intensive irrigation has a

greater effect upon the thermal cycle than the proximity of such large bodies of water as the Aral and even the Caspian Sea (to the north of it).

Of the greatest importance, from the agricultural production angle, is the effect of location characteristics upon the frost hazard. The nocturnal flow off of cold air in calm and clear weather, also, the accumulation of stagnant air in wind-sheltered areas, leads to the formation of frost-bitten parcels. Table 85 cites a typical classification of terrestrial relief by the characteristics of the cold air flow off and the related frost hazard.

TABLE 85

Form of terrestrial relief	Cold air inflow	Cold air flow off	Degree of frost hazard
Summit and upper slopes	None	Present	Minimum
Plains and flat summits	None	None	Medium
Wide open valleys	Mild	Mild	Above medium
Narrow winding valleys	Present	Very mild	Considerable
Synclines	Present	None	Maximum

Height above
sea level

1. Sanatorium X October
2. Valley of Matsesta River
3. High point between the Matsesta
and Agura Rivers
4. Agura River Valley
5. Ascending grade
6. Mountain pass
7. River Khosta Valley
8. Mountain pass
9. SKKO Rest Home
10. Kudopota River Valley
11. Commune imeni Yevdokimov
12. Entrance to Mzypa River Valley
13. Maldanov Bridge
14. Nizhne-Shilovsk Sovkhoz

Figure 57. Thermal survey of the route Sochi-Veseloye, 1934.

A-temperature differentials as observed in January 1934; B -
topographic profile - mean temperature differential derived from
11 series of observations; --- maximum temperature differential as
observed on 9 January.

The reduction of the frost hazard in coastal areas is primarily due to the absence of air stagnancy on account of the prevailing breeze circulation. The moderating effect of bodies of water is particularly pronounced during the end of the summer and during autumn.

For the analysis of the effect of terrestrial relief upon the minimum temperature, the results of the automobile run thermometric surveys (with the aid of photo-thermo-meters) of the route Sochi-Veseloye, 60 kilometers long, conducted under the supervision of G. T. Selyaninov, may be cited. As can be seen from Figure 57 above, the temperature differentials in individual cases were as high as 7 degrees Centigrade.

As was to be expected, the effect of terrestrial relief is at its maximum in clear weather, i.e. in the presence of intensive radiation chilling.

As can be seen from Table 86 below, during non-overcast weather (i.e. when the sum of the total cloud formation for the preceding evening and the following morning is indicated by less than 18 balls), the minimum temperature differentials, as between hill and valley, are about 3 to 4 degrees Centigrade, while under overcast skies, it hardly attains the value of 2 degrees (data taken from observations in Sochi over a hill slope and in Matsesta-over a valley).

[see Table 86 on following page]

The effect of weather conditions upon the minimum tempera-

TABLE 86

Mean minimum temperature differentials Sochi-Staraya
Matsesta for non-overcast and overcast days.

Years	January		February	
	non-overcast	overcast	non-overcast	overcast
1931- 1932	3.3	1.3	2.6	1.0
1932- 1933	3.8	1.3	2.9	1.2
1933- 1934	4.6	1.8	2.5	1.2
1934- 1935	4.3	0.7	2.2	1.2

ture differentials is well illustrated in Figure 58, showing the diurnal variation of the air temperature in Gashnerdi (hill) and Samtredi (valley of the Rion River) on 29-31/XII 1924, in the presence of a cold wave penetration, both curves in the graph run almost tangent to each other, while the weather is cloudy and there is precipitation, with the station in the valley ever registering some excess temperature. However, upon the clearing of the weather, and also upon the abating of the wind, the station in the valley registers a sharp temperature drop.

Since the maximum low temperatures usually occur in the presence of radiation chilling, it becomes clear that the effect of terrestrial relief upon the frost hazard of an area is pronounced

more sharply than its general effect upon minimum temperatures. The variations in the frost hazard are most clearly manifested in the various length of the frost-free interval and in the various value of the absolute and the mean absolute temperature minimums.

The first and the last autumnal freezes are almost always and everywhere accompanied by uniform conditions - clear and calm weather. This circumstance made it possible for I. A. Gol'tsberg to cite a general (for the entire territory of the USSR) characteristic of the location effect upon the length of the frost-free period.

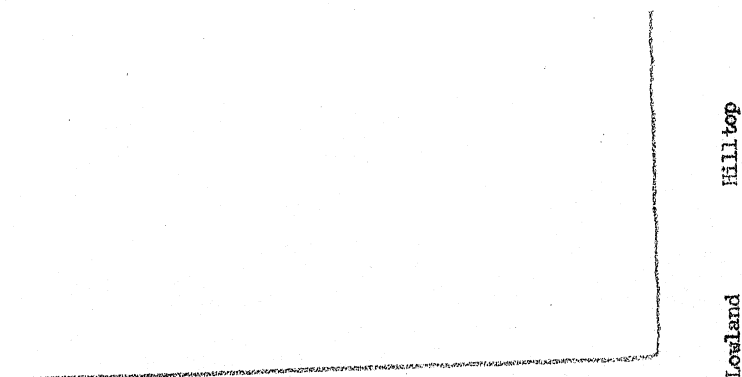


Figure 58. Inversion over Western Georgia.

Table 87 shows (according to I. A. Gol'tsberg) the departures in the length of the frost-free period, under conditions of interlaced relief, from the prevailing length observed at an exposed level area, beyond the direct effect of bodies of water (by the meteorological data collected in the temperate zone of the USSR).

TABLE 87

Variation in the length of the frost-free period in
relation to the location of the meteorological stations

Location	Departure (in number of days) from level area			
	Mean	Maximum	Minimum	Number of Stations
Summits	22	30	15	8
Valleys in a hilly location	-16	-25	-11	22
Mountain valleys-22		-32	-9	16
Synclines and shut-in valleys	-44	-65	-32	9
Islands, sand- bars, littorals	30	44	17	15

In relation to location characteristics, the length of the frost-free period oscillates within the range of 2,5 months. Let us indicate, for purposes of comparison, that the vertical gradient of the frost-free time interval, in its mean value, amounts to 3 to 5 days per 100 meters of altitude.

The effect of bodies of water upon the length of the frost-free period is manifested in a narrow coastal strip only. Even the effect of the Black Sea does not extend beyond a 5 to 7 kilometer -

strip. As to the effect of the Baltic Sea and the large lakes Ladoga and Onega, it is limited to the increase in the length of the frost-free period of the islands and capes only. This is due to the fact that, during the early autumn frosts, usually accompanied by clear and calm weather, there develops a shore breeze, i.e. a surface wind from the land to the water, which neutralizes the direct effect of the water. As a result, the tempering effect of the water is accountable only to the acceleration of turbulent exchange and the abating of the surface temperature inversion in direct proximity to the water line at the shore.

Direct proximity to large rivers also increases the length of the frost-free period. In these cases, the effect of the water basin, as a thermal storage battery, overlaps the effect of the negative form of terrestrial relief.

On the macrogeographic scale, the variegation in the distribution of the length of the frost-free period, is determined by the presence of interlaced terrestrial relief and the continental features of the climate, and is accelerated with the increase in the above characteristics.

Considerable attention to the mean values of the absolute minimum air temperatures, in relation to location characteristics, is given in the subtropical zone, where this particular climatic component is an index for the effectiveness of the growth of a series of valuable agricultural crops: citrus fruits, eucalyptus trees, tung oil trees, and the like.

Because of this, and with the effect of local conditions taken into account, detailed maps of the mean values of the ab-

solute minimum air temperatures are utilized for the purposes of agroclimatic zoning of a territory. A part of such a map constructed by S. A. Sapozhnikora and G. T. Selyaninov, is shown in Figure 59.

above -4°
 from -4 to -6°
 from -6 to -8°
 below -8°

Sukhumi

BLACK SEA

Figure 59. Map of the mean annual values of the absolute minimum temperatures.

On the coastal hills, the mean value of the absolute minimums is above minus 4 degrees. At the coastal lowland - it is from minus 4 to minus 6 degrees, in the valleys - below minus 6 degrees, and, with distance from the sea, it becomes minus 8 degrees Centigrade.

Since the advection factor plays a certain part in the formation of the annual absolute minimums of temperature, the effect of

the bodies of water upon the mean values of the above, particularly under the conditions of Transcaucasia, is manifested with sufficient clarity and is extended rather far, in contradistinction to the effect of the bodies of water upon the early frosts. It being the case that the decrease in the tempering effect of the sea runs, as it does during the summer, in direct ratio to the logarithm of the distance from the water as shown in Figure 60.

Figure 60. The variation in the mean values of the annual absolute minimum temperatures with distance from the sea (plotted by semi-logarithmic coordinates).

1 - Western Trans-caucasia; 2 - Eastern Trans-caucasia. The figures near the dots designate the number of cases.

The relatively large difference in the mean values of the absolute minimum temperatures, in relation to terrestrial relief, is determined by the fact that, in 75 percent of all cases, the absolute minimum temperatures of individual years are accompanied by clear and calm weather, which becomes established after the advection of the cold, i.e., in the presence of radiation chilling, the flowoff of cold air, and the corresponding differentiation of the air temperatures. In the practical utilization of the maps, showing the mean values of the absolute minimum air temperatures, for the evaluation of the frost hazard of an area, it is necessary to remember the following. In 25 percent of cases, the absolute minimums of individual years are observed during overcast and windy weather. The hilltops and the windward slopes are then under less favorable conditions than they are when, in the presence of the

same or even a lower air temperature than the one prevailing in places sheltered from the wind, the chilling of plants, accountable to greater wind velocity, hence, to more rapid turbulent exchange, will occur. Therefore, when selecting parcels for crops highly-susceptible to frost damage, it becomes necessary, in addition to considering the mean value of the absolute minimum temperatures, to also take into account the degree of sheltering from the winds that accompany the cold penetrations.

These remarks on the practical evaluation of the effect of terrestrial relief upon the frost hazard are valid for both the spring and autumn freezeovers and, consequently, also for the length of the frost-free period.

The variation in the mean values of the absolute minimum air temperatures, in relation to location characteristics, for the northwest of the European territory of the USSR, is given by I. A. Gol'tsberg in Table 88 (the plus sign means that the temperature is higher than in an exposed level area, the minus sign - that it is lower).

[see Table 88 on following page]

The above cited data pertain to the climatic evaluation of the frost hazard, and are utilized in assigning crop cultivation to the territories of agricultural economies lying in terrains with interlaced relief. However, the close relation between minimum temperatures and location characteristics must be taken into account also in the current early freezeover forecast. Any further

TABLE 88

Variation in the mean values of the absolute minimum
temperatures of the air in relation to location

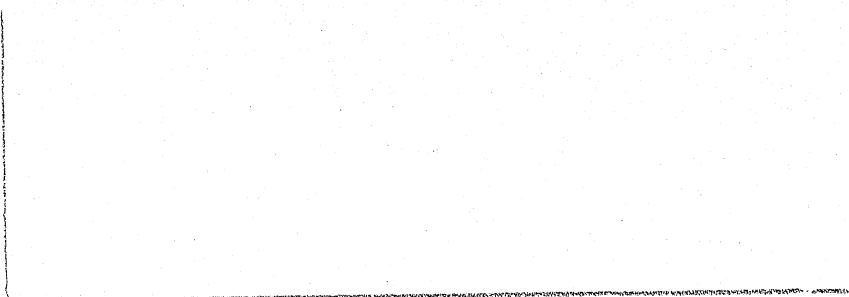
Location	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
Littoral and islands	+2.5	+1.5	+1.0	0.0	+2.0	+2.5	+3.0	+3.5	+3.0	+2.5	+2.5	+2.5
Glades	-2.0	-2.0	-2.0	-2.0	-1.0	-2.0	-2.5	2.5	-2.0	-2.0	-2.5	-2.5
Valleys and lowlands	-3.5	-3.5	-3.5	-3.0	-1.0	-2.0	-4.0	-4.0	-4.0	-3.0	-3.5	-3.5

research in the improvement of forecasting methods must be specifically in the direction of the early freeze forecasts, in view of its importance for agricultural production.

A few short notes on some peculiarities of air humidity, to the extent to which they are determined by the form of terrestrial relief and by the proximity of water basins.

The temperature drop over the negative forms of relief during the night, naturally, causes a rise in relative humidity. A graphic presentation of the temperature, humidity, and wind variation in the pre-morning hours, observed simultaneously at four points disposed on the slope of a small valley, is shown in Figure 61.

meters/second



0650 0700 0710 0720 0730 0740 0750 0800 0810 hours

Figure 61. Simultaneous observations at four points along a valley slope of temperature, relative humidity, and wind. Sovkhoz "Il'-yich, 21/XI 1935, 0645-0835 hours. A - wind velocity; B - relative humidity of the air; V - air temperature; No 1 - H = 0; No 3 - H = 11 meters; No 5 - H = 41 m; No 8 - H = 91 meters.

With the difference in the temperature between the extreme points being 5 to 6 degrees Centigrade, relative humidity varied from 90 to 95 percent at the bottom of the valley, to 50 to 55 percent at the slope at a height of 91 meters. It is of interest to note that, in the presence of foehns and dry winds, the relative humidity in the valleys may come close to 100 percent. Figure 62 is a graphic presentation of the relative humidity and the temperature of the air during dry wind days as observed at the Klonsk Sovkhoz (Western Gruzija) by K. P. Kurskaya - Pakhnevich. While, on the hills, the dry wind began at 0300 hours with a sharp drop in relative humidity to 50 percent, a close to 100 percent relative humidity in the valleys was sustained up to 0700 hours in the morning, when the temperature inversion was cancelled out by the rays of the sun. Even more significant is the rise in relative humidity

to 90 percent in both valleys on the night from the 19th to the 20th of September, while on the hilltops it was only 50 to 60 percent.

0	0600	1200	1800	2400	0600	1200	1800	2400	0600	1200	1800	hours
18/IX				19/IX				20/IX				

Figure 62. Diurnal variation in relative humidity (A) and temperature (B) during dry wind days under various conditions of terrestrial relief. Khousk Tea Sovkhoz, 1935. 1-hill; 2-western slope of transverse ridge; 3- longitudinal valley; 4- sheltered valley; 5- transverse mountain ridge.

On the basis of the same observations, Table 89 cites the recurrence of relative humidity in a longitudinal and transverse (with relation to the wind) valleys, lying in direct proximity to each other, expressed in percentage of its recurrence on a hill.

TABLE 89

Recurrence of relative humidity by gradation during
dry wind periods I/IX - 20/X 1935 (as per Kurzakaya-
Pakhnevich)

Observation Points	Relative humidity, in percent			
	50	40	30	20
Hill	100	100	100	100
Longitudinal valley	63	71	90	33
Transverse valley	46	50	46	0

As was indicated, mountain valley circulation leads to a reverse diurnal variation of relative humidity over the upper part of the slope. According to observations conducted by A. V. Voznesenskiy in the Crimean National Reservation, the nocturnal relative humidity minimum over the slopes of Babugan-Yayla, at an altitude of 1250 meters, is as shown in Table 90 below.

Table 90

Diurnal variation of relative humidity

(mean monthly value, by hygrograph)

Crimean National Reservation (Alabach)

August 1929

Hours

	2100	0200	0400	0600	0800	1000	1200	1400	1600	1800	2000	2200
Relative humidity (in percent)	52	51	52	55	57	58	53	57	59	62	63	56

The proximity of the sea raises relative humidity during the day, with a decisive part played by breeze circulation. Table 91 cites the mean relative humidity at 1300 hours and its mean diurnal value for a series of points in the Rion River Valley, located at various distances from the sea.

Table 91

Mean annual relative humidity (in percent)
at various distances from the sea

Point of observation	Distance from the sea	Mean diurnal value	Mean value at 1300 hours
Poti	Littoral	76	68
Samtredi	60 kilometers	73	63
Sakari	105 kilometers	73	56

Locations, distant from the sea and subject to greater
chilling, may, on the contrary, have higher relative humidities
during the night.

CHAPTER 11

THE EFFECT OF LOCAL CONDITIONS ON LOW CLOUD FORMATION AND PRECIPITATION

As is known, an important part in the formation of clouds is played by the ascending of an air mass, which, in turn, is induced either by convection, or by upslide motion, or by wave motions in the atmosphere. Of particular interest is the development of cumuliiform clouds, stipulated by convection, which, to the greatest extent, is linked to local conditions.

A prerequisite for convective cloud formation is a non-stable atmosphere. In the presence of an inversion layer in the atmosphere, cumuliiform clouds do not develop otherwise than in relation to the characteristics of a locality.

An inversion layer in the atmosphere may be stipulated by local circulation, such as breezes. During the day when, below, the cold and humid air is moving from the sea toward the land, up above there is a reverse and sometimes warmer current. As a result, there is generated, at the boundary of the two currents, a layer of retarded temperature drop, or even of temperature inversion, hindering the formation of cumuliiform clouds. However, at some distance from the sea, at the propagation boundary of the daylight breeze, where strong ascending currents are developed, there occurs considerable cloud formation.

Already A. I. Voyeykov, followed by A. A. Kaminskiy, pointed out that, during the summer, there is less cloud formation and more sunshine over the sea than far away from it. The mean degree of

cloud formation over the islands near the Baltic Sea during the month of June drops to 20 percent, approximating same for Central Asia, while on the coast it is 50 percent, and deeper over the land -- above 55 percent. The low degree of cloud formation over the islands is due to conditions prevailing over the sea, which are non-favorable to convection. The reduced daylight cloud formation on the coast is determined by the above indicated effect of the breeze circulation and the temperature inversion linked to the latter. The reduced cloud formation results in a greater amount of sunshine, exerting a favorable effect upon the climate of the Baltic littoral. This is the reason for the development of a dense network of sanatoria and convalescent homes along the Gulfs of Finland and Riga. The summer diurnal variation in cloud formation over the water is the reverse of same over the land -- the maximum of low cloud formation occurs during the night.

Lakes and large rivers also have a non-favorable effect upon cloud formation in the summer, particularly when their waters are cold, as is the case of the lakes Ladoga and Onega. Cumuliform clouds appear only on the circumference of the water basins. When the wind shifts direction toward the lake, the clouds, carried in its wake, become as if suspended for a while over the mirror-like waters, then rapidly disintegrate, by first decreasing in depth, then in size, until just a haze is visible. No effect of small lakes on cloud formation, as yet, came to notice.

As to large rivers, it is known that, when flying over them, deep clefts can be seen in the uniform cloud banks.

During the cold part of the year, water basins, until such time as they are frozen over, are warmer than the surrounding land. Therefore, there is generated and sustained, over them, a non-stability, resulting in cloud and fog formation. At this time of the year cumuliiform clouds appear over the water basins. However, after they are frozen over, and the snow cover becomes settled, cloud formation over them is no different from that over land.

Without touching upon the peculiarities of cloud formation over mountainous terrain, let us analyze the effect upon cloud formation of elevations and hills. Theoretical computations indicate that even flat elevated plains must induce displacement of the troposphere layers to great heights. Due to orographic convection, low cloud formation over elevations such as the Valdysk and Volga elevated plains is always greater, in its mean value, as compared to that over lower plains, by 10 percent.

The orographic effect manifests itself throughout the year, and, during the cold period, when the air masses are near saturation, it may be even more pronounced. During the cold period of the year, the descent of the lower boundary of cloud formation, which is due to additional condensation, is particularly noticeable.

Local conditions exert an effect upon cloud formation also by way of intermixing of the air masses, which, due to turbulent exchange, are formed at their boundary. This phenomenon occurs over littorals, as, for instance, over the Murmansk littoral.

The variation in cloud formation also affects the amount of precipitation.

A. V. Shipchinskiy, in 1923, after studying the existing precipitation maps of Voronezhskaya Guberniya in juxtaposition to terrestrial relief, points directly to the fact that mean maximum precipitation falls where the terrestrial relief rises from south to north and from southwest to northeast. There is a decrease in the amount of precipitation on the reverse slope.

Later investigations by T. V. Pokrovskaya and A. O. Drozdov confirmed the effect of terrestrial relief upon precipitation. Over the Middle-Russian, the Volyno-Podol'sk, the Donets, and the Volga elevated plains the amount of annual precipitation is by 100-150 millimeters greater than over the surrounding lowlands. Even at such small elevations as the Nevel' Bank and the Siluriysk Plateau in the Leningrad Oblast, an increase in the amount of precipitation can be noticed.

T. V. Pokrovskaya made a special study of the "spottiness" in the distribution of precipitation, and came up with two causes of maximum precipitation: (1) the additional condensation of water vapors, due to the local intensification of vertical currents, and (2) the redistribution of precipitation by the wind.

The first cause reduces itself to greater cloud formation, already mentioned above, the second one ties in the distribution of precipitation with the velocity of the wind. Falling raindrops and snow particles are subject to the effect of two forces -- the force of gravity and the force of the wind. With an acceleration in the wind over the windward hill slopes, the resultant of these two forces obtains a greater horizontal component, and precipitation

is carried off by the wind. In those spots where the wind becomes abated, the force of gravity prevails, and precipitation falls vertically to the surface. Thus, in a locality with interlaced relief, over parcels, where there is accelerated wind velocity, a diminished amount of precipitation is to be expected, while over parcels with decelerated wind velocity, an increased amount of precipitation can be anticipated.

If, as in the case of rain, the above indicated trends may be debatable, there is certainly no doubt, as to their validity, in the case of snow, in which the differentiation is additionally accelerated by the subsequent blowoff of the snow from the windward slopes into spots that are sheltered from the wind (see Chapter 11). Since the peculiar characteristics of the distribution of precipitation over a given territory are determined by the direction of the air current, prevailing directly at the time precipitation is falling, the summary effect increases in areas where a definite direction of the wind during precipitation predominates. Having specified more accurately the effect of terrestrial relief upon the falling of precipitation, and knowing the "rainy wind-rose", it will become possible to interpolate the results of observations at individual points upon their surrounding territory with greater veracity than is being done now.

Large bodies of water also exert an effect upon the distribution of precipitation. There is less summer precipitation on islands and flat littorals. At the propagation boundary of the day breeze, where strong ascending air currents are evolved, the amount of precipitation is usually increased. As an example, the

precipitation maximums along the north shore of the Gulf of Finland, as represented in Fig 63, can be cited. The effect of the breeze is reinforced here by the effect of the hills, the sea-facing slopes of which are windward in relation to the prevailing winds.

GULF OF FINLAND

Lake Ladoga

Fig 63. Distribution of annual amounts of precipitation (in millimeters) along the shores of the Gulf of Finland (as per Yefimova).

No decrease in the amount of precipitation is observed over elevated windward shores. For example, in Lake Baykal, the mean annual amount of precipitation over the island does not exceed 150 millimeters, and tuberous vegetation is growing there, while on the windward southeast shore, literally pressed in against the mountains, the amount of annual precipitation is in excess of 500 millimeters.

CHAPTER 15

THE CLIMATE OF A FOREST

In a dense forest, with the tree tops joined, the active surface coincides with the surface formed by the tree tops. In such a forest, solar radiation practically does not penetrate to the surface of the soil. The basic surface of evaporation is also the surface of the tree tops. The same tree tops reduce the wind velocity to a minimum, practically to a complete calm.

In the case of a sparse forest, the situation is somewhat different. The active surface, in some part, in relation to the degree of rarefaction, is adapted to the soil and to the grass cover. Still, the velocity of the wind is abated. As a result, specific conditions of thermal economy are created, whereby a rarefied forest, by the thermal cycle, approximates the conditions of steppe or field parcels, and has little in common with a dense forest. Some analogy may be drawn between sylvan and herbaceous vegetation. The thermal cycle of a field with a sparse grass cover differs little from that of denuded parcels.

In the formation of the climate of the forest, an important part is played by the species composition of the forest, the presence of trees of the second stratum and the young growth of trees. In addition, the climate of the forest is in direct relation to zonal climatic conditions, also to the nature of terrestrial relief and of the soil. It is to be regretted that all these

factors are not taken into account to a sufficient degree, as a result of which contradictory conclusions are drawn by individual authors.

Nevertheless, some characteristics of the climate of a forest can be considered as established, if not quantitatively, at least qualitatively.

Let us analyze the factors which determine the heat exchange in a forest.

The radiation cycle was already discussed in Chapter 1. Solar radiation is almost completely intercepted by the tree tops. The same tree tops also protect the surface of the soil, the moss cover and the grass cover from loss of heat through effective radiation. In a dense joined forest, the loss of heat through long-wave radiation occurs exclusively from the surface of the tree tops.

An indirect characteristic of turbulent exchange is the velocity of the wind. Fig 64 is a graphic presentation of the variation in the wind velocity with height in two types of a pine forest: without young growth and with trees of the second stratum and young growth. Up to the top boundary of the tree tops the wind velocity changes little with height, but above the tree tops, it increases in jumps.

(For Figure 64, see following page)

The velocity of the wind in a sparse forest is, in all cases, twice as great as that in a dense forest. Figure 65 shows a vertical

profile of wind velocity in an oak forest, foliated and denuded. When the velocity of the wind over the forest is 4 meters/second, its velocity in a foliated forest is 1 meter/second or less, in a denuded forest, 1.5-2.0 meters/second, and only at the very surface it is decreased to 1 meter/second.

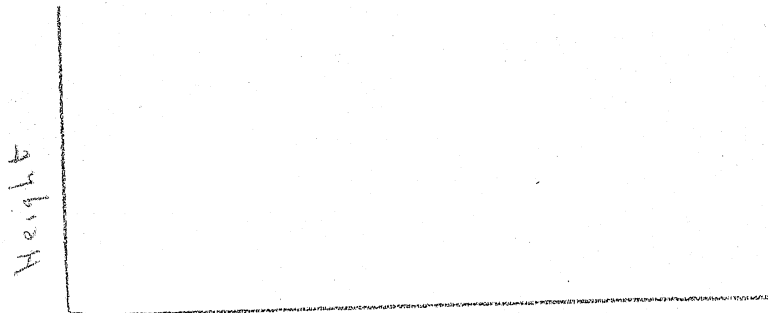


Figure 64. Variation in wind velocity with height in a pine forest. I -- forest without young growth; II -- forest with second stratum and young growth.

In a sparse forest without young growth, there occurs sometimes, under the tree tops, in the midst of the tree trunks, an acceleration in the wind velocity, and one velocity minimum is adapted to the soil surface, another one -- to the tree tops.

If we compare the velocity of the wind in a deciduous forest at the height of 2 meters with its velocity in an exposed level place, we discover that in sparse forest, during the winter (i.e., without foliage), it will be 40-60 percent, and, during the summer, it will be 30-40 percent of the wind velocity in an exposed place.

In the case of a dense forest, the figures will be, respectively, 20-30 percent in the summer.

The abating of turbulent exchange in a forest is on the same order. The tree tops sort of isolate the depth of the forest from the free atmosphere. During the day, the abatement of turbulent exchange under the joined tree tops, as compared to an exposed place, must be even greater, due to the absence of a superadiabatic temperature gradient (see further in the text) and its accompanying thermal convection.

Above the tree tops, turbulent exchange, on the contrary, is accelerated. The vorticity in the air current over a forest is easily discernible during low-altitude flights by the jerks sustained by the plane.

A. A. Shepelvskiy (by the observations of the Pavlovsk Observatory at a height of 45 meters over the ground) came to the conclusion that winds blowing from the park (northerlies and westerlies) are distinguished by greater turbulence, as compared to the winds blowing from the side, where there is no forest (southerlies).

Meters

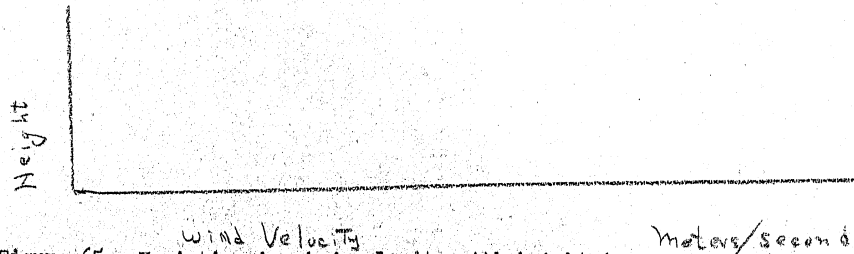


Figure 65. Variation in wind velocity with height in an oak forest, foliated (I) and denuded (II).

At the edge of the forest, in clear calm weather, "forest winds", similar to breezes, may be evolved. According to G. N. Vysotskiy, the cool air over the shaded edge of the forest, during the day, flows in a direction from the forest to the field, while over the sunlit edge of the forest, the warm air rises upward.

The peculiarity of the radiation cycle and of turbulent exchange lead, in a dense forest, to a vertical temperature distribution, which is substantially different from that over an exposed level place. During the night, there is the most frequent occurrence of isothermy, a mildly pronounced minimum is adapted to the upper part of the tree tops, which is, essentially, the active surface. The leveling off of the temperature is helped along by the downflow, from the tree tops, of the cooled and heavier air. During the day, a temperature inversion is observed, and the temperature maximum is adapted to the same upper part of the tree tops.

Figure 66 is a graphic presentation of the temperature variation at given heights (3, 11, 19, 23 and 27 meters), in an oak forest, at sunrise and at near-meridional hours, on a clear summer day (as per Geyger and Amann). Observations were made with the aid of relatively heavy electric thermometers (thermocouples). These are subject to considerable radiation error, therefore, readings correspond not as much to the temperature of the air as to the temperature of the thin branches, leaves, etc.

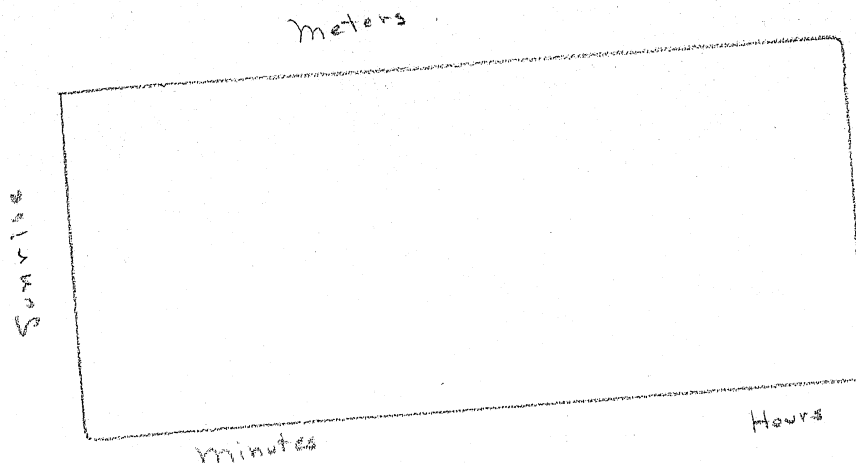


Figure 66. Temperature variation at various heights in an oak forest (by Geyger and Amann). I -- at sunrise; II -- during the day; 1-27 meters (over the tree tops); 2-23 meters (in the tree tops); 3-19 meters (under the tree tops); 4-11 meters (in the young growth); 5-3 meters (over the ground).

At sunrise, i.e., while the radiation balance still has a negative value, the vertical temperature variation is small. (within the limit of one degree Centigrade), with the minimum in the upper part of the tree tops. During the near-meridional hours, the vertical temperature variation is rather considerable -- up to 9 degrees Centigrade. The maximum is observed at the height of 23 meters, i.e., in the tree tops, the minimum -- at the height of 3 meters over the ground. The air temperature, of course, has no such vertical difference, but the general character of the temperature inversion remains.

Observations by V. N. Obolenskiy in Lesnoye (Leningrad) on air temperatures in the midst of young fir trees and young oak trees (under conditions of clear weather), confirm the active part

played by the surface of the tree tops (Table 92).

Table 92

Air temperatures in the midst
of forest plantings

Hours	Months	Species of Trees	Temperature			
			At the surface of the soil	Inside the tree tops	At the tips of tree tops	Above the tree tops
1300	VI	Young fir	19.7	18.8	23.2	20.5
	VII	Young oak	19.2	20.1	22.1	21.6
2100	VI	Young fir	12.5	12.2	11.8	13.6
	VII	Young oak	14.2	13.1	12.8	15.0

In the midst of young fir, as well as in the midst of young oak, the maximum and minimum air temperatures are adapted to the upper part of the tree tops. During the night, the vertical temperature variations are relatively small (0.7-1.4 degrees), while during the day the temperature inversion in the layer from the surface of the soil to the tree tops measures 3-4 degrees.

In all those cases, when the tree tops shade the soil, the daylight temperature inversion observed in the forest results in the maximum temperature of the air layer up to 2 meters high being lower than in an exposed place, while, during the night, due to

isothermy, it is warmer in the forest, i.e., the minimum temperature is higher. This peculiar characteristic of the forest is confirmed by considerable empirical data, including the data collected by A. P. Tol'skiy, relating to pine tree plantings of the Buzuluk forest (Kuybyshev Oblast).

Table 93 is a comparison of observations in a forest with observations in a large glade, since no data was available for an exposed place. It is to be remembered that the differences in the temperature as between forest and glade are usually greater than between forest and field.

Table 93

Oscillation limits of the air temperature
differences (forest-glade) during individual
hours (at the height of 2 meters), Buzuluk

Pine Forest, 1904-1910

Temperature Differentials	Semi-annual period	
	Cold	Warm
Mean monthly	from -0.17 to 0.00	from +0.40 to +0.62
Mean maximum values	from -1.08 to -0.72	from +0.70 to +0.92
Mean minimum values	from 0.68 to 1.02	from 0.42 to 0.77

The character of the maximum and minimum temperature differences remains the same through the entire year, but the mean monthly temperature differential changes. During the summer season, when

the number of daylight cycle hours predominates, the forest, in the mean diurnal value, is colder than the glade. During the winter, the characteristics of the nocturnal and daylight cycles are in mutual equilibrium, as a result of which the mean monthly temperatures are close to each other in value.

In nearing the surface of the soil, the air temperature differentials, as between a dense forest and an exposed place, become increased at the expense of the difference in the vertical temperature gradients, particularly on summer days, when there is a temperature inversion in the forest, and a superadiabatic gradient in an open place. In these cases, the temperature of the surface air layers in a forest is ~~by~~ several degrees lower than in an exposed place.

The reduction in the maximum and the increase in the minimum temperatures result in a diminished diurnal temperature range in a forest.

With the increase in minimum temperatures is connected an increase in the length of the frost-free period. It is known that the young frost-sensitive fir trees perish from the frost in exposed places and develop normally under the canopy of a forest.

An even greater effect is exerted by the forest vegetation upon the temperature of the soil. During the warm period when the soil is warmed through, the temperature of the soil in a forest is considerably below that in an exposed location. According to A. P. Tol'skiy (Table 94), the temperature of the upper metric layer

of the soil in a young forest of pine-tree plantings is by 5 degrees Centigrade lower than the same in a glade.

Table 94
Mean monthly soil temperature differentials
(forest-glade) from April to September,
Buzuluk pine forest, 1904-1910

Depth in centimeters	IV	V	VI	VII	VIII	IX
Surface	-3.5	-5.1	-5.7	-6.5	-4.8	-3.3
10	-3.4	-6.0	-6.3	-6.8	-4.6	-2.5
25	-3.6	-6.0	-6.9	-7.4	-5.1	-2.7
50	-2.2	-7.3	-7.0	-7.1	-5.4	-3.2
100	-0.9	-6.0	-6.2	-6.5	-5.4	-3.9
200	-0.2	-3.6	-4.8	-5.3	-4.9	-4.1

For purposes of comparison, let us indicate that, on a macro-scale, under the same conditions, such differentials in summer soil temperatures occur, for instance, between Vologda and Poltava.

The diurnal soil temperature range is also considerably smaller in a forest. According to A. P. Tol'skiy, on the sand dunes of the Buzuluk Forest, untouched by hewing, the diurnal temperature range on the surface of the soil was only 50-60 percent of the temperature range on the soil surface of a large glade.

Juxtaposing the diurnal variation in the heat exchange of the soil over the summer season, A. P. Tol'skiy concludes that the heat

exchange in the soil of a forest equals 55-60 percent of that in a glade. The correlationship of the respective annual thermal economies is approximately the same (± 1.0 great cal/square centimeter for a forest, ± 1.5 great cal/square centimeter for a glade).

During the winter, the soil temperature is higher in the forest, as shown by Table 95.

Table 95

Mean monthly soil temperature differentials
(forest-glade) from October to March,
Buzuluk pine Forest, 1904-1910.

Depth in centimeters	X	XI	XII	I	II	III
Surface	-1.5	0.1	0.8	0.7	0.9	0.1
10	-0.2	0.5	1.0	0.6	0.6	-0.4
25	0.1	1.1	1.0	0.5	0.1	0.0
50	-0.3	1.3	1.3	1.3	0.9	0.5
100	-1.7	0.1	0.5	0.6	0.4	0.4
200	-2.7	-1.1	-0.2	-0.3	-0.1	-0.1

The forest litter (fallen leaves and coniferous needles, the moss and grass covers) play an important part in the thermal economy and, together with that, in the thermal cycle of the forest soil. Hindering the warming of the soil during the summer, it also prevents it from chilling and freezing in the winter. The snow cover, which,

in most cases, is deeper and less compact in the forest than outside of it, also protects the forest soil from winter chilling and freezing.

The above indicated peculiarities of the thermal cycle pertain to dense forests only. In a sparse forest, conditions may be substantially different, since some diminution in the intensity of the radiation balance, at the expense of the partial shading by the tree tops, is, to a certain degree, compensated by abated turbulent exchange. As a result of abated heat exchange with the upper air layers, the daylight air temperature in such a forest may be even higher than over an exposed location.

In turning to the problem of the moisture economy, it is first of all necessary to analyze the precipitation cycle.

Over a series of decades, considerable attention was given to the study of the effect of forests upon precipitation, particularly in dry areas. The difficulty in solving this problem is aggravated by the fact that the pluviographic readings in the field and in the forest are non-comparable, particularly in the winter. Strong winds usually carry away the snow from the pluviometers installed in exposed areas, while in the forest the pluviometers are protected by the abated velocity of the wind.

In addition, the precipitation itself that is falling over the field and the forest is not amenable to comparative study, since part of it, falling over the forest, does not reach the ground and, in being detained by the tree tops, is evaporated directly from their surface.

Observations conducted in the Buzuluk pine forest show that the pine tree tops detain 12 percent of precipitation in the form of snow and 28 percent in the form of rain.

Pine tree plantings detain over the annual period only 12-14 percent, and medium-growth fir trees detain 36 percent of precipitation. As the amount of precipitation increases, the percentage of it detained by the tree tops is decreased.

Some increase in the amount of precipitation, particularly in the winter, in the presence of low cloud formations, is attributable to the forced rising of the air current over the forest. But this amount cannot be considerable, since the height of the forests in dry areas rarely attains 20-25 meters.

With reference to the increase of precipitation over the forest, which is attributable to the additional humidification of the atmosphere, due to the accelerated evaporation from forested parcels, this, too, cannot be of any great importance, since the additional evaporation, as compared to evaporation from herbaceous vegetation, proceeds at the expense of a diminution in the flowoff, which, in the contemplated dry areas, amounts to only 15-20 percent and less of the annual precipitation total. Besides, the evaporated moisture does not remain in place, but is carried away by the air currents into other areas.

Of great importance, particularly in some areas, may be the so-called "horizontal precipitations", settling down in fog on the trees in the form of hoar-frost and fog-drops ("namoros").

A. D. Zamorskiy defined "namoros'" as water in liquid state precipitated from fog.

The precipitation of water from fog is of particular importance at the edge of a forest, at the windward slope, and at the mountain top. The quantitative characteristics of this phenomenon are circumscribed by hoar-frost, i.e., fog precipitation in a solid state.

According to A. Lyutnitskiy, a snow cover, 10 centimeters in depth, may be formed in a forest by hoar-frost. I. Nikitin described the formation of a sled tract in a forest from crumbling hoar-frost.

According to observations by G. N. Vysotskiy in the Mariupol' Forest Reservation, the amount of collapsing hoar-frost in each individual case is 0.7 to 1.6 millimeters.

If we accept this minimum figure of 0.7 millimeter, and the mean annual number of days with hoar-frost as 60, we derive 35 millimeters, which is 9 percent of the annual precipitation total.

A. D. Zamorskiy observed several times a fog-drop rain in the forest at a time when not one meteorological station, located near the forest, or even inside the forest, registered any perceptible precipitation. A denuded deciduous forest, even far away from its edge, produces a considerable amount of fog-drop precipitation. A fully foliated deciduous forest detains the fog moisture more rapidly, causing the amount of fog-drop precipitation to drop sharply

in the direction from the edge of the forest into its depth.

The part played by hoar-frost and fog-drop precipitation is particularly great in mountainous areas, where frequent fogs prevail. The quantitative study of these "horizontal precipitations" is an urgent problem not to be delayed.

Outside their effect upon the amount of precipitation, the forest plantings exert a substantial effect upon the snow cover, its depth, consolidation, and duration.

The basic reason for the greater depth of the snow cover in the forest, as compared to the field, is the absence of a wind sufficiently strong to deteriorate it. The depth of the snow cover is particularly great at the edges of the forest, due to the additional drifting of the snow from the adjacent exposed areas. In addition, the above mentioned "horizontal precipitations" add to its depth, with a greater amount of said "horizontal precipitation" distributed to the forest edge. However, the detention of the snow by the tree tops, particularly the coniferous tree tops, inversely, tends to diminish the depth of the snow cover, since the snow evaporates more rapidly from the tree tops. This circumstance sometimes causes the snow cover in the depth of a dense coniferous forest to be of lower depth than in the field.

Investigations by I. S. Nesterov at the Petrovskaya "Dacha" of the Timiryazev Agricultural Academy (Moscow), as far back as the beginning of the century, have shown that the snow water reserves in various forest plantings may differ substantially (Table 96).

Table 96

Moisture reserves in the snow cover
according to I. S. Nesterov, March 1906

Character of forest plantings	Moisture reserves in millimeters
Young saplings, 2-4 years old, in small glades and forest clearings	129
Birch planting 35-75 years old	128
Oak planting 25 and 90 years old	141
Oak planting 25-30 years old	78
Pine-tree planting 60-90 years old	81
Fir-tree planting 25-35 years old	54
Pine-tree planting with birch admixture 65-75 years old	111
Same with admixture of larch	79
Same with admixture of fir	73
Experimental field	79

The maximum snow water reserves are accumulated in deciduous forest plantings, in pure coniferous plantings they are considerably smaller, by 40 and even 60 percent. The admixture of birch considerably increases the water reserves.

According to data assembled by A. P. Tol'skiy in the Borovoye reservation of the Buzuluk pine forest, the snow cover in the forest persists for one week longer than in the field. As per observations in the Parfinskoye reservation (near Storaya Russa) and at the Petrovskaya "dacha" near Moscow, i.e. under more northerly conditions, the snow is retained in the forest for longer periods - by 2 weeks or even a month, as compared to the field.

The lag in the melting of the snow cover in the forest, which is of great importance in the control of the water flowoff, is due not only to its greater depth, but also to the abating of the turbulent exchange, which plays a leading part in the melting of snow at the expense of the advection of the warm humid air. The melting of the forest snow attributable to radiation heat may proceed even somewhat more intensively. The albedo of the tree trunks, which is considerably below that of the snow, causes an increase in the absorption of solar radiation, due to which the snow around the trees melts already during the first days of sunshine, forming the well known "craters".

The greater depth of the snow cover and its slower rate of melting contribute to the greater accumulation of moisture in the forest soil as compared to that of the field, particularly in the presence of the forest litter (fallen leaves and coniferous needles, grass and moss covers).

N. F. Savykin points out that the soil under the litter, even in freezing, remains porous and water-permeable. Contributing, as it does, to the percolation of the melt and rain waters into the depth of the soil, the forest litter has a beneficent effect upon the water economy, by decreasing the surface flow-off. This is the predominant water-preservation function of the forest.

According to N. F. Savykin, the rate of water absorption by the soil, in the presence of forest litter, is accelerated three-fold, the indicated acceleration varying, however, in relation to the nature of the litter. The highest and the most rapid rate of water absorption is manifested by a moss, particularly a sphagnum cover. The deterioration of the litter, as, for example, by grazing cattle, decreases the rate of water absorption by the soil, reducing thereby the water-preservation role of the forest.

The greater moisture reserves in the soil may contribute to some increase in the evaporation from the forest as a whole, compared to the evaporation from adjacent fields. Without considering the forest edges, in which the moisture reserves are particularly great at the expense of the adjacent exposed parcels, the increase in the moisture reserve of the forest proper may be attributable to the diminished flow-off, which under conditions of a plain, is rarely in excess of 20 percent of the annual precipitation total (and under arid conditions, not more than 10-15 percent) and to the "horizontal precipitations", which may constitute 10 percent of the precipitation total.

Evaporation from a forest (including transpiration by the plants and evaporation from the soil as well as from the tree surfaces) is determined by the character of the forest planting, general climatic conditions, and the peculiarities of location.

The wide-leaf deciduous forests of Belorussia and North-western Ukraine, where the annual precipitation total is in excess of 600 millimeters, have the capacity and actually evaporate considerably greater amounts than the "tayga" in Yakutia with an annual precipitation total of less than 300 millimeters. The haloxylon forests in the desert area of central Asia evaporate particularly small amounts of moisture. Eucalyptus trees on the contrary evaporate great quantities of moisture, and they are specially cultivated for the drainage of swamps.

Forests, situated in the bottomlands of rivers and at the places of egress of ground waters, evaporate more than forests situated in watersheds with deep ground-water level. Their evaporation is limited not by the amount of precipitation, but by the underflow of ground waters and in the presence of an unlimited inflow of the latter, by the thermal energy requisite to evaporation.

During the summer radiation heat provides in the forest and steppe areas, for a diurnal evaporation rate, of approximately 5 millimeters. A greater evaporation rate of the great forest bulks is only possible at the expense of heat advection, as becomes available in the presence of dry winds.

The potential evaporation capacities of a forest on the whole, including the trees at all levels, the undergrowth and the

grass cover, may be greater than those of the grass cover in a meadow or field, particularly at the expense of the longer vegetation period of the forest. Therefore, a forest may render the soil arid to a greater degree than steppe parcels, but only in the case, when the evaporation surplus will be in excess of the moisture income at the expense of a diminished flow-off and additional precipitation. The problem of the agro-forestry melioration experts consists in developing such types of forest plantings, as will promote an increase in the moisture income, yet, will not run out of bounds in the expenditure of the latter.

Research, conducted by I. M. Labunskiy in Veliko-Anadol', effectively confirmed the part played by the selection of the proper varieties of trees in the cultivation of forests in accordance with climatic conditions. The so-called arboreal-umbral type of plantings, with oak as the main species, and pointed-leaf maple and linden to size, situated at watersheds, turned out to be not only the most adaptable, in the sense of survival, under steppe conditions, but also capable of basic benevolent changes, to the level of the ground waters and, together with that, to the moisture content in the soil of the adjacent fields.

In planting forests in semi-arid areas, it is necessary to take into consideration the climatic possibilities or adapt the forest plantings to parcels with an underflow of surface or ground waters, as for instance, to the drops in the ground known as "padiny". The "padiny" in the Caspian lowland are drops in the terrestrial relief, to which the water flows during the melting of the snow. In such localities, the growth of the planted forest

will be provided for not only by local precipitation, but also by precipitation falling upon adjacent areas.

Although, on the whole, a forest may evaporate no less than a field or a meadow, yet, under the canopy of the forest, due to the shelter from the sun and to the abated turbulent exchange, the evaporability and even the evaporation itself is considerably less intensive as compared to a field. According to A. P. Tol'skiy, evaporability of a forest, as registered with the aid of a Wild's evaporimeter, over the period from May to September, is only 40-50 percent of the evaporability of a glade. The greater degree of dampness of the surface layers of forest soil, as compared to the soil of a field or a wasteland, shows that in a forest, there is less evaporation directly from the soil.

As to the air humidity and its vertical variation in a forest, it is primarily determined by the evaporation from the tree tops, by the abated turbulent exchange inside the forest, and by the temperature cycle.

In a dense forest, the maximum absolute humidity during the daylight hours is adapted to the tree tops, as was observed by V. N. Obolenskiy in a young fir tree forest. Absolute humidity is diminished in an up and down direction from the tree tops. The maximum relative humidity in a dense forest is observed more frequently above the tree tops and near the very soil, since it is adapted to the setting in of the temperature minimum.

The comparison of the absolute humidity at the height of 2 meters in a forest and over a field, as per observations in the

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The comparison of the absolute humidity at the height of 2 meters in a forest and over a field, as per observations in the

Borovoye Experimental Forestry Reservation, showed the absence of any perceptible differences. Relative humidity at the same height is considerably higher in the forest (by 2-4 percent) than in the field.

In generalizing all the above said with reference to the microclimate of the soil and its adjacent air layer in a dense forest, we come to the conclusion that it is substantially different from the microclimate of a meadow and field. Abated illumination, a uniform temperature variation (with small diurnal oscillations, with longer frost-free periods, but also with lower daylight temperatures), a mild wind, and low evaporability create altogether unique conditions for processes taking place in the soil, as well as for the growth of the grass and moss covers. It is no mere accident that, specifically under the canopy of the forest, the tundra vegetation penetrates far to the south. And there is small wonder that after a forest is cleared away, a change in the soil and grass cover occurs. Plants, that are shade-loving and not adapted to sudden temperature changes and high evaporability, are gradually displaced by the vegetation of the meadow and the steppe.

The basic features of the climate of the forest, i.e., subdued illumination, a more uniform temperature cycle, diminished wind velocity, and low evaporability, persist all the way to the upper surface of the tree tops. All this makes the conditions of the development of a tree inside the forest, on its edge, and in the open field very different, respectively.

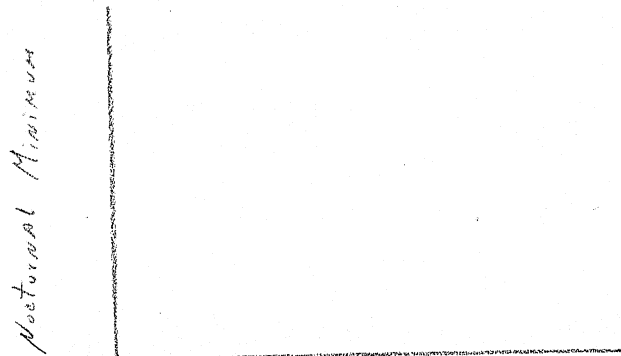
Particular conditions are also created in forest glades.

The climate of forest glades is in relation to their size and the characteristics of the forest edge. Small glades, shaded in by surrounding trees, which are not isolated from the forest by dense edges of undergrowth and bush, are little different in their climatic cycle from a sparse forest, but with the increase in the area of the glades and the extent, to which they are surrounded by dense forest edges, the climate of the glades is changed abruptly.

In small glades, not separated by dense forest edges, the income of solar radiation as well as effective radiation, is diminished at the expense of the darkening caused by the surrounding trees. This draws them together with the forest. The unhampered exchange between such a glade and the forest levels off the emerging differences. In the case of large glades, the darkening by surrounding trees plays an insignificant part. But the presence, near the glade, of a dense forest edge hinders the exchange between the forest and the glade. The diminished wind velocity (as compared to same over an exposed place) weakens the turbulent exchange over the glade. In the aggregate, all the above brings to large glades, having dense forest edges nearby, unique conditions, more different from the forest than the open field. The climate of glades has much in common with the climate of valleys and even the climate of syndines.

First of all, in such glades, the diurnal temperature range rises sharply. This is attributable to the higher daylight temperatures and, particularly, to lower temperature minimums.

Figure 67 is a graphic presentation of the increase in the frost hazard with the increase in the diameter of the cradle-clearing of the forest. With the diameter of the clearing increased from 20 to 80 meters, the minimum temperature during 17 cold days dropped from 7.5 to 4 degrees Centigrade. I. A. Gol'tsberg characterizes the diurnal temperature range and the mean value of the absolute temperature minima in terms of departures from the values of the same elements prevailing in an exposed level location (Table 97).



Diameter of cradle-clearing

Area of cradle-clearing

Ratio of diameter of clearing to height of forest

Figure 67. The increase in frost-hazard with increase in the area of the cradle-clearings of forests (according to Geyger)

1 - observation period of 17 cold spring days; 2 - 6 June 1940

TABLE 97

Departures of the mean diurnal air temperature range
A and the mean value of the absolute air
temperature minima M over glade

	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
A	0.5	1.9	1.0	0.0	0.5	0.5	1.0	1.0	1.0	0.5	0.0	0.5
M	-2.0	-2.0	-2.0	-2.0	-1.0	-2.0	-2.5	-2.5	-2.0	-2.0	-2.5	-2.5

The length of the frost-free period, in the case of a glade, is diminished by 20-30 days, as compared to an exposed location. Thus, the climate of a glade turns out to be more severe than the climate of an open field, not to mention the climate of a forest. The diminution of wind velocity promotes a more uniform distribution of snow by comparison with an exposed place, and also a reduced evaporability. Terrestrial relief (valley, slopes) may intensify or weaken the peculiar features in the climate of glades.

CHAPTER 16

THE CLIMATE OF A CITY

The climate of a city is a complex phenomenon, determined not only by the character of the buildings, architectural development, vegetation and landscape layouts, but also by the terrestrial relief of the locality involved. An important part is played by the presence of industrial plants.

Special chapters are devoted to the effect of terrestrial relief and vegetation. We will, therefore, analyze in this chapter only the effect of architectural development and the effect of industrial plants upon the climate of a city.

Gaseous and solid industrial wastes (smoke, soot), together with the dust in the air, form a film extending over the city. In London, a mean value of more than 10 grams of soot and other air contaminating impurities precipitates upon a surface of one square meter over a period of one month. This figure may be doubled over industrial centers.

The contamination of the atmosphere diminishes its transparency, resulting in a diminution of direct insolation and also effective radiation.

In Leningrad, according to G. A. Lynboslavskiy, the mean intensity of solar radiation is by 17 percent lower than in Pavlovsk, which lies only 25 kilometers to the south. This difference is in relation to the direction of the wind, and when the latter blows from the sea, the difference is reduced to 10 percent. However, the diminution of direct insolation is in

some part compensated by an increase in diffused radiation and a decrease in radiation loss of heat by effective emission. As a result, the daylight radiation balance, even if it changes, it does so only to an insignificant degree. The nocturnal radiation chilling diminishes to a more substantial degree.

In the presence of a snow cover, the city precipitation of soot and smoke promote the soiling of its surface, thereby reducing the value of its albedo and causing it to disintegrate by one-two weeks ahead of the snow in the city environs.

The typical active surface in a city is constituted by the iron roofs, the stone walls of buildings, the asphalt and cobblestone of the street pavements. Evaporation from these surfaces, and, consequently, loss of heat to evaporation is insignificant since precipitation without being detained runs off by the way of gutters, drain pipes and sewerage system.

The iron of the roofs in addition has a low thermal capacity, the heat conduction of iron and stone being relatively high.

Wind velocity (see Table 70) and, consequently, turbulent exchange over a city are abated. All this, in the aggregate, raises the temperature of a city by comparison with its environs, particularly during the evening hours, when the buildings, heated up during the day, gradually give off their heat into the air. Table 98 cites the data collected by L. S. Berg, for Moscow, Leningrad and their environs, relating to the mean air temperatures.

TABLE 98
MEAN MONTHLY AIR TEMPERATURES

Name of observation Points	Height above sea level (in meters)	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
Moscow-city (former Land Survey Institute)	167	-10.2	-8.5	-4.1	4.1	12.7	16.4	18.7	16.5	10.8	4.4	-2.3	-7.5
Environs of Moscow (Petrovsko-Razumovskoye)	160	-10.8	-9.1	-4.8	3.4	11.8	15.6	18.0	15.8	10.1	3.7	-2.8	-8.0
Differential		0.6	0.6	0.7	0.7	0.9	0.8	0.7	0.7	0.7	0.7	0.5	0.5
Leningrad-city (Vasil'yevskiy Island)	6	-7.6	-7.7	-4.1	2.8	9.5	14.6	17.5	15.5	10.6	4.7	-0.9	-5.5
Environs of Leningrad (Forestry Institute)	26	-8.1	-8.3	-4.7	2.5	9.4	14.2	16.8	14.7	9.8	4.0	-1.4	-7.0
Differential		0.5	0.6	0.6	0.3	0.1	0.4	0.7	0.8	0.8	0.7	0.5	1.5

In the mean value, it is warmer in the city, by more than 0.5 degrees Centigrade, than in the environs.

The absolute temperature minimums are also higher in the city. For the period 1910-1926, the mean value derived from the absolute annual temperature minima for Moscow was minus 28.0 degrees Centigrade, for Petrovsko-Razumovskoye - minus 31.2 degrees Centigrade. During individual years, the differential in the absolute temperature minima was as high as 6 degrees Centigrade. In Leningrad, at the Vasilyevskiy Island (Central Physical Observatory), the temperature on 25 December 1892 descended to minus 34 degrees Centigrade, while in Lesnoye (the environs of the city), the reading was minus 41 degrees Centigrade. Some lower temperature minima differentials, as between a city and its environs are cited by A. A. Kaminskiy, for Khar'kov, Voronezh, and Penza, which is due to a lesser architectural development. It was noted that damage, sustained by vegetation through early freezeovers, was less frequent within city limits.

Higher temperatures in a city, by comparison with its environs, increase with the growth of the city and its architectural development. This is manifested by the century-wise temperature cycle of the air. In all large cities (Leningrad, Berlin and others), the temperature is slowly rising through a century, beyond its relation to the centurial cycle, determined by macro-processes.

The temperature cycle is different in different parts of a city, in relation to the extent of green areas, the width of the streets, the material of the buildings.

Observations made on clear calm days show that particularly intense heating effects are sustained during the day by wide paved streets and plazas. Boulevards and plazas overgrown with trees heat up to a much lesser degree. Narrow lanes shaded by tall buildings are the coolest spots during the day.

The absolute, as well as the relative, humidity in a large city is somewhat lower due to abated evaporation. However, the sprinkling of the streets raises the humidity.

The wind over a large city undergoes substantial deformation. In well built-up streets, the wind is predominantly blowing along the street in either direction. Its direction may shift rapidly without coinciding with the general direction of the air current over the city. Vertices evolve readily over streets and street crossings. The vorticity structure of the city wind is plainly visible during snowstorms, when on one side of the street the snow is blown off at the pavement, while on the other side the wind piles it into large drifts. Streams of wind, shifting transversely across the streets, form a sort of closed circulation, with the ascending branch along one side of the street, and the descending one - along the other side.

Figure 68, showing the results of an anemometric survey, gives some idea on the nature of the surface air currents. In the presence of a basic air current with a velocity of 7 meters/second, the wind velocity over the streets and lanes, lying along the direction of the wind, the wind velocity was 4 meters/second over the lanes, disposed perpendicular to the direction of the wind,

a mild longitudinal movement of the air, with constantly shifting direction, is observed. Over narrow passes, the wind velocity may increase.

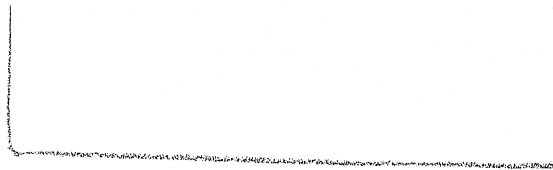


Figure 68. Air currents over a city, in the presence of a north-easterly wind.

When city areas are overheated, convection currents are evolved over them, together with a system of mild winds from the periphery toward the center. Individual circulations are also evolving over individual streets. Along the sunlit walls the air moves upward, along the shaded walls - downward. In nocturnal calm, the air downflow occurs along both sides of the streets, and the upflow over its center.

The products of combustion constitute excellent condensation nuclei, by the virtue of which the occurrence of fogs over a city is increased.

Table 99 cites the number of days with fog over Moscow and over one of the population points of the Moscow Oblast, far away from the industrial centers (data collected by A. I. Danilin).

TABLE 99

Recurrence of fog

Years	Moscow	Sasovo
1930	109	39
1931	160	39
1932	173	33
1933	144	72

Moscow was not always subject to its present degree of fog recurrence. Figure 69 shows a graphic presentation of the fog cycle over the 35-year period (1900-1934) over Moscow. It illustrates well the increase in the recurrence of fog, which, of course, is attributable to industrial growth.

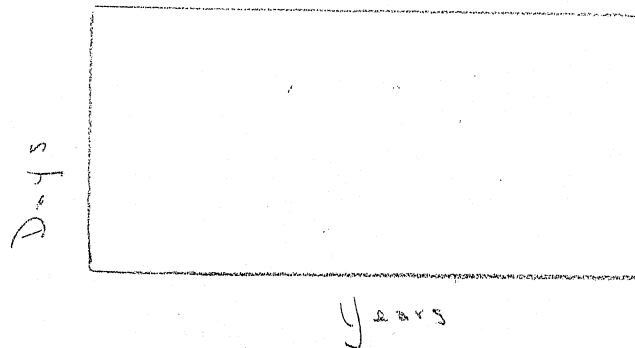


Figure 69. Fog cycle over Moscow for the 35-year period (1900-1934)
(according to A. A. Danilin).

The presence of condensation nuclei, and also convection, may result in the formation over a city of cumuliiform clouds, which, in turn, reduces the number of clear days. There is data pointing

to an increase in precipitation over large cities.

During the winter temperature inversions over industrial areas, throwing up great amounts of smoke into the atmosphere, there is frequent recurrence of the so-called "inversion haze", with its greatest density at the upper inversion boundary. The inversion haze, considerably lowering visibility, may be sustained for weeks, and becomes diffused only with the deterioration of the inversion.

In eastern Siberia, due to a sharply pronounced temperature inversion, such hazes and fogs evolve over relatively sparsely populated locations, and they are intensified with the accelerated use of heating furnaces.

Temperature inversions, in weakening turbulent exchange, promote the stagnation in the surface air layers of poisonous industrial wastes, which phenomenon in some cases, as in the Ruhr Valley, is known to have resulted in mass toxic effects.

The toxic effects, entailed by the proximity of industrial plants to population centers, is taken into consideration in the urban planning and zoning of the USSR. The industrial and plant-site part of a city is disposed from the direction of the winds of least recurrence. In addition, it is isolated from the housing areas by a protective zone of gardens and parks. Landscaping and vegetation planting is done in the areas of the plants themselves, resulting in a modified daylight heating effect, and also reducing the dust content in the air. Housing projects are also extensively enlivened by landscaping and greenery.

In the southern areas, the planting of trees is of particular importance, since the trees, when coming into full growth, shade the streets from the direct sun and soften the effect of the daylight sultriness. The radiation cycle and, together with that, daylight heating are controlled by the orientation and the width of the city blocks.

SECTION V

THE STRUGGLE AGAINST THE DESTRUCTIVE PHENOMENA OF
CLIMATE AND WEATHER IN THE SURFACE LAYER OF
AIR AND IN THE SOIL

As was already pointed out in the introduction, micro-climate and local climate are of particular interest to us, since changing them in the proper direction holds great possibilities, and in a series of cases it has already entered into the practice of our national economy.

One of the oldest methods of changing the local climate and microclimate - irrigation, has an effect upon the water and thermal cycles of the surface air layers and the soil. The features of the latter were analyzed at the end of Chapter 8 and partly in Chapter 13. The change of the thermal cycle in the presence of irrigation can be considered as an accompanying phenomenon. The basic purpose of irrigation is, as it is known, the melioration of the water cycle of the soil.

In this section, we will familiarize ourselves, by way of individual examples, with the basic principles of the thermal melioration of the climate of the soil and the surface air layers, of the struggle against late autumn or early spring freezeovers, of the meteorological effectiveness of the field-shielding forest strips and snow detention. It being the case that the industrial effectiveness of the above, inasmuch as it is determined by a series of economic factors, is not within the scope of this work, and will not be gone into.

The greatest effect, of course, is produced by complex

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The greatest effect, of course, is produced by complex

measures embracing, if possible, all the components of the thermal balance. Under all conditions, it is necessary to take into account the general weather and climatic factors, since the latter determine to a considerable degree the effectiveness of the measures undertaken. For example, a change effected in the radiation characteristics of the active surface, justifies itself only in the presence of a well pronounced radiation balance, i.e., in clear weather.

CHAPTER 17

THERMAL MELIORATION. SHIELDED GROUND

Thermal melioration of the soil in the northern areas consists in raising its temperature, in the southern areas, in a series of cases, on the contrary, - in lowering its temperature.

The control of the thermal cycle of the soil may proceed by way of changing the radiational characteristics of the active surface, changing the losses of heat to evaporation, changing the turbulent heat exchange soil-air with the aid of insulation covering (mulching), and, finally, by the way of the so-called sealed off ground, when not only the soil, but also the air layers, contiguous to it, are insulated from the free atmosphere (hotbeds and hothouses).

Raising the temperature of the soil by way of changing its radiational characteristics, as, for instance, by blackening, produces the greatest effect in the case of light-colored denuded soils with considerable albedo values (20 percent and over). In this case, the blackening of the soil, let us say, with coal powder, which lowers the albedo value down to 5 percent, will increase the absorption of direct insolation by 15 percent and more. The presence of vegetation, of course, will reduce the radiational effect of the blackening of the soil. This effect is also small during overcast weather.

In order to lower the temperature of the soil, its albedo is increased by painting it white. Thus, the whitening of the trunks and branches of fruit trees reduces their temperature,

which, in turn causes a lag of 4-5 days in the blossoming and inflorescence, thereby reducing the probability of damage from spring freezeovers.

A great effect upon the climate of the soil can be attained by mulching, i.e., by covering the soil with special paper (mulch), straw, bitumen, glass. In mulching, the soil is covered in such a way as not to interfere with the growth of agricultural crops. Mulching, in all cases, sharply reduces the evaporation from the soil surface. The thermal effect, however, is various. The covering of the soil with bituminous emulsion raises its temperature by an average of 2 degrees. Straw, on the contrary, during the warm period, reduces the temperature of the soil.

Glass frames produce a particularly great effect. In covering the soil directly, they raise its temperature by 5-7 degrees, which is due not only to a favorable radiation cycle, in the form of a sharp reduction in effective emission, but also by a diminution in the loss of heat to evaporation.

Under natural conditions, the role of mulch, which, in the summer, retards the heating of the soil, is played by sod and any vegetation cover present. Thus, in accordance with observations, conducted by P. I. Koloskov in the Far East, the "culturizing" of the soil, i.e., the removal of the sod and plowing, leads to the modification in the permafrost level, particularly under conditions of simultaneous measures for the increase in snow accumulation, i.e., reduction of winter heat emission.

The analysis of the effect of the grass cover upon the temperature of the soil was given in Chapter 9. Agrotechnical measures controlling the depth of the grass cover and its compactness (the denseness of the young crop, fertilizer) determine the peculiarities of the thermal cycle of the soil. Investigations of interest in this field were conducted by A. M. Shul'gin in Nal'chik. He observed the temperature of the soil under winter wheat: (1) continuous seeding with 15 centimeters between seedbeds and 1.5-2 centimeter seed distances within the beds; (2) with 70 centimeters between seedbeds and 3-centimeter seed distances within the beds; (3) with 70 centimeters between seedbeds and 12-centimeter seed distances within the beds. For purposes of comparison, he also observed the temperature under black fallow.

Results of these observations over a 5-day period are given in Table 100. The lowest temperatures were observed under the continuous seedbeds of wheat, the highest temperatures were under the black fallow. Wide-bed seeding, with

a feed area of 70 x 3 centimeters, are, by the temperature of the soil, close to continuous seeding, while wide row seeding with a feed area of 70 x 12 centimeters approximate the temperature of black fallow.

A considerable variation in soil temperature occurs also under millet with various feeding areas (30 x 1.5 and 60 x 5 centimeters). The difference in soil temperature between these parcels during the phase of inflorescence and ripening, at 1300 hours, on most of the days, was from 6 to 10 degrees Centigrade at the depth of 3 centimeters, and 3 - 0 degrees at the depth of 10 centimeters. Observations on winter wheat and millet indicate that the decisive effect upon the soil temperature is exerted by the amount of vegetation mass per unit of area and the degree of its density.

The greater the vegetation mass, the lesser the daylight warming of the soil, the smaller the entire diurnal soil temperature range.

P. I. Koloskov proposes the following empirical formula for the diurnal temperature range, linking the temperature range of a soil, covered with a grass cover (A_T), with the temperature range of denuded soil (A_0):

$$A_T = A_0 - \frac{\sqrt{m}}{k} \quad , \quad (25)$$

where m is the weight of air-dry vegetation mass in grams per one square meter; k is coefficient, the value of which approximates 2.

Thus, substantial changes can be made in the thermal cycle

of the soil by controlling its vegetation cover.

In order to meliorate the thermal cycle of the soil and the contiguous air layers, the control of the heat exchange soil-air must be utilized. The abating of wind velocity and, by the same token, of the heat exchange in the surface air layers during the day and, inversely, the increase in the heat exchange during the nocturnal hours promote an increase not only in the temperature of the soil, but also in the temperature of the contiguous air layers during the day at the expense of a reduced heat emission into the free atmosphere, and at night at the expense of the heat increase from the free atmosphere.

The echelon formations of the tall herbaceous and undergrowth vegetation decrease the turbulent exchange not only during the day, but also at night. Nonetheless, they raise the mean value of the thermal cycle of the soil and the adjacent air layers, since during the spring and summer the daylight cycle is of decisive value in the general thermal economy.

[See next page for Table 100]

M. V. Shokhin used for the control of the heat exchange between the soil and the air, in the Main Botanical Garden of the Academy of Sciences, USSR, gauze slideways in the form of panels 2 meters long and 1 meter wide. During the day, the gauze slideways shielded the plots of land under agricultural crops from the wind, and for the night they were rolled up in order not to hinder the turbulent exchange. The mean maximum temperature among the gauze slideways, as compared to the control plot, was by 2.6 degrees

TABLE 100

Temperature of the soil at the depths of 3 and 10 centimeters, under winter wheat of various sowing techniques and under fallow, at 1300 hours, observed at Nal'chik, 1939

Date of Observation	3 Centimeters				10 Centimeters			
	Continuous Sowing	70 x 3 cm	70 x 12 cm	Fallow	Continuous Sowing	70 x 3 cm	70 x 12 cm	Fallow
June								
28	26	26	32	33	22	22	24	26
29	26	30	34	38	22	22	24	28
30	28	28	34	38	22	22	24	28
July								
1	28	28	32	34	22	22	24	27
2	26	28	32	36	23	23	24	28

higher at the height of 5 centimeters, by 1.9 degrees higher at the height of 20 centimeters, and by 0.4 degrees higher at the height of 40 centimeters.

The accelerated accumulation of heat during the daylight hours was also conducive to a certain increase in the minimum temperatures (by 0.8 degrees on the surface of the soil). As a result, the melon crop amid the gauze slide-panels turned out considerably higher by comparison with the control parcel.

During the winter, the basic method in raising the thermal cycle of the soil and in meliorating the conditions for the hibernation of the winter crops is snow detention. In addition to snow detention, accomplished with the aid of field-sheltering forest strips, to which a special chapter will be devoted, great effects can be accomplished by the method, proposed by Academician T. D. Lysenko in 1941, of sowing winter crops along the path of the non-treated stubble of the summer crops. This method of sowing solves the problem of the insulcation of winter wheat into the Siberian steppes, which are distinguished by severe cold and inadequate snowfall.

Lysenko established the fact that the loss of winter wheat is due not to the direct effect of low temperatures, but to the mechanical injuries sustained by the above-surface and subsurface parts of the plants. The above-the-surface parts of plants, when not covered with snow, become fragile in the presence of severe frost, break and crack in the strong winds, while the subsurface parts are torn up by large ice crystals forming when the turned over friable soil freezes through. When the method of sowing the

winter crop along the stubble path of the summer crop is used, conditions for the hibernation of winter crops are sharply improved. The stubble in itself reduces the velocity of the wind by increasing the roughness of the underlying surface, thereby preventing mechanical injuries to the winter crop. From the moment of snow precipitation, the stubble promotes a uniform, deep, and permanent snow cover, which protects the winter crop both from mechanical injury and from the effects of the low temperatures. In Table 101, A. Golovlev cites comparative data for the depth of the snow cover and temperature at the depth of the fasciculation node (about 3 centimeters), in relation to the method of sowing.

TABLE 101

Depth of snow cover and soil temperature at the depth of
the fasciculation node of winter crops -- Barnaul Selective
Station, 1947 - 1948

Character of Soil	10/XII	20/XII	30/XII	10/I	20/I	30/I
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Soil Temperature

Along stubble	-2.5	-14.5	-11.2	-4.2	-4.5	-4.9
Along fallow	-8.5	-20.5	-10.4	-6.4	-8.4	-12.2

Snow Cover (in Centimeters)

Along stubble	25	18	34	32	32	36
Along fallow	14	10	10	12	23	30

According to A. Golovlev, winter crops, when sown in fallow, perish to the extent of 82 - 100 percent, while, when sown in the

stubble remaining from the summer crops, only from 1.5 to 3.5 percent of the crop is lost.

The greater depth of the snow cover promotes greater dampness of the soil, compensating thereby the autumnal lack of moisture in the stubble as compared to fallow.

Snow detention can also be effected with the aid of slide-panels. In this case, the part of these slide panels is played by the rows of annual plants (as sunflower), sown all around the fields allotted to winter crops. Such slide-panels may attain 1 meter or even more in height. By reducing the velocity of the wind in the surface air layer, they thereby promote snow detention.

Table 102 cites data by A. M. Shul'gin indicating the high effectiveness of this method. The slide panels detain even the first snow. As a result, by the end of December, when the severe frosts set in, the snow cover attains a depth of 30 - 40 centimeters, while in the fields, where no snow detention was undertaken, its depth amounts to only 10 - 17 centimeters. This differential in the snow cover depth remains during the second half of the winter. The slide-panels retain their importance to the very end of the winter, when a considerable amount of snow is still retained by them for an additional 1 - 2 weeks, while it is already gone from the open fields, exposing vegetation to the vicissitudes of abrupt temperature fluctuations. The freezing of the soil in the open plots in the Barnaul area penetrated to a depth of 1.5 - 2.0 meters, while in the plots, where the snow was detained, the frost penetrated to a depth of 50 - 70 centimeters only.

TABLE 102

Snow cover on winter crop fields, with and without
snow-detaining slide-panels (in centimeters) -

Barnaul, 1941 - 1948

Type of Soil	XI	XII	I	II	III
With slide-panels	20	38	45	52	53
Without slide-panels	8	15	17	17	12

The effectiveness of the slide-panels is determined by the distance between them. As per A. M. Shul'gin, the most effective snow-detaining job is done by slide-panels located on 3.6-meter centers.

Close-set slide-panels are particularly effective during winters with strong winds. Data collected by A. M. Shul'gin, as presented in Table 103, show the effect upon the snow cover of slide-panels with various center distances for the windy winter of 1942 - 1943.

The increase in the depth of the snow cover, naturally, raises the temperature of the soil, particularly the minimum temperatures due to the leveling of oscillations. Table 104 (as per A. M. Shul'gin) cites the data on the minimum soil temperatures at the depth of the fasciculation node of winter crops (3 centimeters) for fields without and with snow detention.

The minimum temperatures amid the slide-panels were always considerably above those prevailing in the plots without the slide-

panels, it being the case that, in the severest winter of 1942 - 1943, these differentials were at their maximum attaining a value of over 20 degrees Centigrade. The difference in the thermal cycle manifested itself on the status of the winter crops: the percentage of loss in the plots with snow detention hardly reached 15, while in the plots without snow detention it fluctuated between 60 and 100 percent.

Accumulating a mighty snow cover, the slide-panels, naturally, increase the moisture content in the soil, as can be seen from Table 105 (as per A. M. Shul'gin).

TABLE 105

Depth of the snow cover in fields with slide-panels placed on various centers (distances between panels in centimeters), Barnaul, 1942 - 1943.

Type of Field	23/XI	29/XII	20/I	10/II	26/II
Without slide-panels	6	4	3	10	24
Slide-panels every					
10.8 meters	8	9	11	13	31
Same every 7.2 meters	14	15	16	27	45
Same every 3.6 meters	19	23	30	36	60

TABLE 104

Effect of snow detention on the minimum temperatures
of the soil at the depth of 3 centimeters

Years	Type of Field	Depth of Snow Cover (cm)	Minimum Temperature
1941/42	With panels	30-58	-14.0
	Without panels	0-30	-26.2
1942/43	With panels	20-60	- 8.0
	Without panels	0-24	-32.0
1943/44	With panels	35-51	-15.0
	Without panels	0-14	-23.1
1944/45	With panels	31-50	-12.9
	Without panels	0-20	-18.9

TABLE 105

Effect of snow detention upon the accumulation of active
moisture in the 1-meter soil layer (in millimeters) -
Barnaul

Type of Field	1943 - 44			1944 - 45		
	Autumn	Spring	Δ	Autumn	Spring	Δ
With panels	115	180	65	95	139	44
Without panels	116	131	15	98	105	7

In fields without snow detention, the increase in moisture reserves, attributable to winter precipitation, is negligible (only up to 7 - 15 millimeters). Snow detention with the aid of slide-panels increases it 4 - 6 times over.

There are other methods of snow detention, it being the case that all of them are evaluated not merely by the meteorological effect, but also from the angle of agrotechnical requirements, into the details of which we cannot enter here. The examples already cited well illustrate the possibilities pertaining to the melioration of the climate of the soil, its thermal and moisture cycles, which are opened up with the control of the snow cover.

In conclusion, let us dwell upon the climate of the so-called "shielded soil", in which classification belong hotbeds and hothouses of various design. The problem of shielded soil assumes ever greater importance on a national economic scale.

The beneficial effect of the hotbed is manifested with particular effect in the presence of late autumn and early spring freezovers. During the September frosts the temperature of the soil surface in the hotbed was by 5 - 6 degrees Centigrade higher than in the exposed ground (see Table 106).

See next page for Table 106

TABLE 106

Temperature of the soil in exposed ground and in hothouse under glass cover

Month and observation Point	Soil Surface				Depth of 10 Centimeters			
	0700 hours	1300 hours	2100 hours	Mean	0700 hours	1300 hours	2100 hours	Mean
August								
Exposed ground	17.6	25.8	14.6	19.3	16.5	19.6	19.0	18.4
Hothouse	19.8	29.5	17.9	22.4	19.0	21.5	21.5	20.6
September								
Exposed ground	9.2	16.0	9.0	11.4	11.8	12.3	12.1	12.1
Hothouse	12.6	21.1	12.7	15.4	14.2	15.2	15.6	15.0
October								
Exposed ground	6.2	9.8	5.9	7.3	7.5	8.3	8.2	8.0
Hothouse	8.1	12.8	8.3	9.7	9.6	10.1	10.2	10.0

The beneficial effect of the hothouse was particularly pronounced during the out-of-season freezeovers. In the September frosts, the soil surface temperature was by 5 - 6 degrees Centigrade higher in the hothouse than in the exposed ground.

I. A. Bayakovskiy investigated the meteorological cycle of a solar hothouse (without artificial heating) near Sverdlovsk in 1932. Over the period 21/VI - 20/X, the mean temperature of the air in the hothouse was by 3.6 degrees Centigrade higher than outside. The mean temperature maxima for the same period differed by 6.5 degrees, and the minima -- by 2.5 degrees. On the days with

freezovers, which, over the investigated period, numbered 44, the mean value of the minimum temperatures, as between the hothouse and the outside, was 5.5 degrees, with only six cases of the temperature descending to below zero in the hothouse. Hothouses are also distinguished by greater air humidity, both absolute and relative (by the mean value of 10 percent).

In hothouses with supplementary artificial heating, any thermal cycle may be established. In special winter houses, even artificial illumination is provided. Their particular characteristics are already beyond the range of microclimatic research.

Chapter 18

COPING WITH OUT-OF-SEASON FREEZEOVERS

Under the conditions of continental climate prevailing over the greater part of the Soviet Union, the problem of coping with the out-of-season freezeovers is of exceptional importance.

During the spring and autumn seasons, in the central and southern parts of the USSR, and during the winter, in the subtropical area, due to great diurnal temperature ranges, the daylight temperatures provide for the normal growth of vegetation, while the nocturnal freezeovers tend to destroy vegetation within a period of several hours. The struggle against frost, therefore, becomes instrumental in the fullest utilization of the beneficial features of the climate.

At the northern agricultural boundary, protection from the frost is necessary for potatoes and grain cereals, in the central belt -- for fruit trees, in the south -- for cotton, and in the subtropical belt -- for tangerines, oranges, and other subtropical crops.

The protection of fruit trees and vineyards from the pernicious effect of freezeovers by the method of ~~smoke-curing~~ is known for the longest time.

In addition to ~~smoke-curing~~, special heating devices are employed. Experiments in temperature-raising during freezeovers were also attempted by stirring the air with the aid of propellers

and by raising the heat emission from the soil. Various types of soil covering are also employed. These methods are usually based on the peculiarities of the weather conditions accompanying the freezeovers.

Although in most cases the freezeovers are evolved against the background of cold-air advection, the decisive part is played by the nocturnal chilling at the expense of the intensive effective radiation. The temperature inversion, forming under these conditions, weakens the turbulent exchange, i.e., prevents the arrival of warmth from the upper air layers, not yet chilled, and by virtue of this, additionally promotes the chilling of the surface air layer and the active surface itself, including the surfaces of vegetation, their foliage, flowers, and fruits.

The liberation of heat during the condensation of water vapor, particularly during the formation of dew, in the case when the dew point is above zero Centigrade, hinders the formation of frost in the same manner as does the good thermal conduction of the soil in the case of its adequate warming through during the daylight hours. Thus, it can be seen, that all the components of the thermal balance play their parts in the evolution of freezeovers.

In the subsequent analysis of the measures undertaken to cope with the freezeovers, we will proceed from the angle of controlling each individual freezeover.

Low wind velocities, which are one of the reasons for the sharp drop in the nocturnal temperature, are at the same time instrumental in the execution of most of the countermeasures, which

are usually of a local nature, and their success, therefore, tied in with low mobility of the air.

The effectiveness of the counterfrost measures is determined by the differential in the air temperatures as between the heated parcel and the check parcel. Since the minimum temperature, under natural conditions, is greatly variable, its variation may be one of the causes of contradictory evaluations. In addition, the thermal effect of the countermeasures depends on the peculiarities of the local climate. The degree of self-containment of the local climate, the part it plays in the formation of local advection -- the flowoff and inflow of cold air,^a circumstance overlooked up to a recent date, is of great importance.

The utilization of the differential between the air temperatures of the heated parcel and the check parcel, as the index for the effectiveness of the counterfrost measures, may sometimes result in an apparent minimization of its practical importance. In the final analysis, it is not the temperature of the air that is so important, but the temperature of the vegetation itself, which, under natural conditions, is lower than the air temperature during the hours of nocturnal chilling. Therefore, the countermeasures will produce a practical effect under the conditions of raising the temperature of the plant itself, even in the presence of only a small temperature differential between the air of the heated parcel and the air of the check parcel.

It is only natural that the most effective counterfrost methods are the complex ones, changing not one, but several components of the thermal balance.

Let us now analyze the oldest of all methods -- ~~smoke-curing~~. The theoretical analysis of this method was made by M. E. Berlyand and P. N. Krasikov. In the presence of the smoke screen, the diffusion of nocturnal chilling occurs basically under the effect of three factors:

- (1) the diminution of effective emission due to the presence of the smoke;
- (2) the direct propagation of heat from the source of the smoke, since the formation of smoke is always accompanied by the liberation of heat;
- (3) the liberation of heat during the condensation of vapor on the hygroscopical smoke particles.

According to M. E. Berlyand, the dilution of effective emission ΔR by the smoke-curing may be presented approximately by the following formula

$$\Delta R = R \left(1 - e^{-\alpha \frac{p}{u}} \right) \quad (26)$$

where R is the effective emission at the check point, in the absence of curing; p is the consumption of smoke-producing substance, u is the velocity of the wind, α is the coefficient of absorption, the value of which equals 1500 - 3000. Formula (26) shows that the diminution in effective radiation ΔR is the greater, the greater the value of R itself. In addition, the value of ΔR is increased

with an increase in the consumption of the smoke-producing material and with a decrease in wind velocity.

The thermal effect of the diminution in effective emission, i.e., the surpassing of the air temperature over the smoke-cured parcel, as compared to that of the check parcel, at the expense of the diminution in effective emission, is determined not only by the diminution of effective emission at the given moment, but by the length of the smoke-curing period, by the velocity of the wind, by the turbulent exchange, by the thermal characteristics of the soil, and by the height over the active surface.

The thermal effect of smoke-curing increases in direct ratio to the diminution in effective emission, and is approximately in proportion to the square root of the smoke-curing time. The thermal effect is substantially abated with height. The increase in wind velocity and in turbulent exchange, also in the heat conduction of the active surface, reduce the thermal effect of the diminution in effective emission, since, in these cases, even without the smoke-curing, the chilling of the surface air layer is relatively small, due to the intensive inflow of warmth from the soil and from the air. Due to this circumstance, the thermal effect of the smoke-curing of a parcel, covered with a thick grass cover, hindering the flow of heat from the soil, will be greater as compared to that of a parcel covered with a sparse grass cover, particularly in the presence of damp soil. The thermal effect of smoke-curing will be greater over peat-soil parcels than over mineralized soil parcels.

When smoke-curing with "shashki" [logs] or smoke-piles, a

certain amount of heat is liberated in the burning process. According to P. N. Krasikov, 700 great calories are liberated by the combustion of one smoke "shaska", weighing 1.8 kilograms, in 10 minutes of burning time, and in the burning of a conventional smoke pile, 1000 great calories, per each kilogram weight of the pile, in 2 - 3 hours of burning time, are liberated.

Due to this, the air, surrounding the source of the smoke, will become warmed, with the thermal effect approximately in inverse ratio to the distance from the smoke source, to the velocity of the wind, and to the coefficient of turbulent exchange. The warming effect is also decreased with height.

As to the liberation of heat during the condensation of water vapor it can only take place when the smoke contains hygroscopic substances, and when the humidity of the air is adequately high. The condensation of the water vapors, in addition to the direct thermal effect, promotes the diminution of effective emission.

Experiments, conducted in the Central Geophysical Observatory by M. E. Berlyand and P. N. Krasikov, showed that the simultaneous burning of 50 white smoke "shashki" on a parcel of 1 hectare, in the presence of a wind velocity of 1 meter/second at the height of 2 meters and a relative humidity of about 90 percent (the burning time for 1 "shashka" is 10 minutes), creates a thermal effect equal to about 1 degree Centigrade, with the parts played by the three above indicated components approximately equal. In the presence of a relative humidity below 80 percent, the thermal effect is reduced, since the heat of condensation, in this case, can be considered equal to zero.

The rise in temperature is substantially in relation to the velocity of the wind. In calm weather, the burning of the same number of "shashki" will raise the temperature by 2.5 degrees Centigrade. Inversely, with the velocity of the wind being 2 meters/second, the thermal effect will not exceed 0.5 of a degree. The thermal effect of smoke piles is of a lower value. It takes the burning of a 100 one-kilogram pile on a parcel of 1 hectare, in the presence of a 1 meter/second wind, to raise the temperature by 1 degree Centigrade.

The most effective method of coping with the out-of-season frozecoovers, at the present time, is considered to be the open-air warming of the plants. In this method, the warming takes place at the expense of the heat developed by the burning of solid or liquid fuel in special heaters. The heat is propagated from the heater both through turbulence and radiation. According to V. S. Lavriychuk, who conducted his experiments on the lemon plantations of the subtropical zone of Western Transcaucasia, 100 two-pipe petroleum heaters, operating over a 1-hectare parcel, warm up the air by 2 - 2.5 degrees Centigrade, and 400 heaters -- by 5 degrees Centigrade.

The Nikiforov heater is the best in existence. It consists of a petroleum tank-reservoir and a cylindrical pipe 1.5 meters high. In relation to the width of the pipe, this heater consumes 1.2 - 1.75 kilograms of fuel per hour.

The maximum thermal effect should be derived by the combination method of such heating and smoke-curing. It is to be regretted that, as yet, the open-air heating method is used only in

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The maximum thermal effect should be derived by the combination method of such heating and smoke-curing. It is to be regretted that, as yet, the open-air heating method is used only in

the subtropical zone. This is due partly to its relatively high cost, making its application feasible in the case of the most valuable crops only.

In the presence of artificial irrigation, the fields are sprayed prior to the anticipated freezeovers.

In Central Asia, the spraying of the cotton fields prior to the anticipated freezeover raises the temperature by 2 and even 3 degrees Centigrade. This is due to two factors: the liberation of heat in condensation on account of a higher dew point, and an increase in the heat emission from the soil, the heat conduction of which increases with greater moisture.

Research, conducted by R. I. Aseykin and others in the Physico-Agronomical Institute, showed the effectiveness of the so-called aeration method in coping with the freezeovers, particularly in the case of potato plantings in the northern areas. The potatoes are planted on 25 - 30 centimeter-high ridges, with wide spacings between rows and large distances between the bushes in the rows. In the presence of such a distribution, the daylight losses of heat to evaporation-transpiration, per area unit of active surface, are reduced, which results in greater daylight warming of the soil. During the night, the soil emits this heat into the air, which circulates freely in the midst of the sparse grass cover, compensating for the heat lost by the latter through effective emission.

The practice of covering the plants for the night or for even longer periods of time is also to be counted among the methods of coping with the out-of-season freezeovers. In the subtropical

belt, lemon trees are kept under gauze covers throughout the entire cold period. This method was developed by G. B. Nadaraya, who demonstrated its complete effectiveness. Such protective jackets are effective, not only against radiational, but also against advective freezeovers.

Chapter 19

THE EFFECT OF FOREST STRIPS UPON THE
CLIMATE OF THE ADJACENT FIELDS

The first experiments in forestation, under conditions of the steppe, for the protection of the fields from the pernicious effects of the droughts and dry winds, that were made on the territory of the USSR, go back to the beginning of the 19th century. Yet, only the catastrophic drought of 1891, with the terrifying hunger in its wake, forced the tsarist government to turn its attention to forestation in the drought-plagued steppe areas.

In 1892, a special expedition was organized under the direction of V. V. Dokuchayev, with an assigned task to test out and list "various methods and procedures relating to the forest and water economies in the steppe areas". At the watershed, dividing the Dnepr, the Donets, the Don, and the Volga, 3 experimental parcels were laid out: the Kamenmostepnoy (Khrenovsk), Starobel'sk, and Veliko-Anadol'sk (known since 1889 as the Marinpole Experimental Forest Range).

The research conducted by this expedition, which was headed by talented scientists, brought out the basic characteristics of the effect of forests and forest strips upon the local climate and microclimate of the fields and upon the increase in the yield of agricultural crops. However, in 1899 the Dokuchayev Expedition was liquidated, the experiments, initiated by it, gradually folded up,

and only under the Soviet power, beginning with 1923 - 1924, new parcels of forestation strips have appeared in the various areas of the arid zone. Of particular interest are the observations organized by the All-Union Agro-forest-melioration Research Institute at the Timashev parcel of the Kuybyshev Oblast, at the Gusel'sk and Rostashev parcels of the Saratov Oblast, and at Bogdin parcel of the Stalingrad Oblast. The work of these experimental stations is summarized in the reports of G. I. Matyakin and others.

In addition, a study of the effect of field-shielding forestation strips, with the meteorological environment follow-up, was conducted in the Ukraine by the Ukraine Research Institute for Agro-Forest Melioration and Forest Economy at the Vladimirov Experimental Station of the Nikolayev Oblast, at the Bogdanovka sovkhos in the Odessa Oblast, and in a series of other localities.

The experiments initiated by V. V. Dokuchayev in the Kamennaya steppe were successfully expanded by the Agricultural Research Institute of the Central Black Earth Belt imeni V. V. Dokuchayev.

All these investigations, and also the direct production experience of the kolkhozes and sovkhoses in the Rostov and Stalingrad Oblast's, were made the foundation of a decree by the Party and the Government pertaining to the plan on field-shielding forestation strips and other measures for coping with droughts and dry winds, published on 20 October 1948.

In accordance with this decree, over the period from 1949 to

1965, there are to be brought into existence field-shielding forestation strips to cover an area of 4,172,500 hectares, exclusive of those to be planted by the State (118,000 hectares), and exclusive of forestation plantings in valleys, ravines, sand dunes, etc. (1,858,500 hectares).

The decree also specifies the widths of the forestation strips. For the protection of the fields from dry winds under conditions of a plain, the forestation strips are to be 10 - 20 meters wide. Under conditions of interlaced relief, the forest strips, 20 - 60 meters wide, are to be planted across the slopes, to stop the erosion and the washing away of the soil. The width of forestation strips around bodies of water is to be 10 - 20 meters. In ravines and valleys, in addition to strips, continuous forestation is to be instituted.

The shielding forestation strips are to be disposed along the fields in a formation of cells stretched out across slopes and athwart the path of the dry winds. The form and the areas of the inter-strip terrain are to be such as to be amenable to mechanized cultivation with the aid of complex agricultural machinery. Therefore, the length of an inter-forestation strip field is to be above 1000 meters, while its width, depending on climatic conditions, is to be from 200 to 500 meters. The effect of the climatic conditions is manifested not only directly, but also indirectly, since they determine, to a certain degree, the height of the forestation strips, with the exception of irrigated plots, where irrigation overlaps the role of the climate, and the trees may attain their natural ultimate heights.

The economic effectiveness of forestation field-shielding strips is determined by a series of factors, the most important one being the hydrometeorological factor.

The hydrological factor is manifested in the sharp diminution of the surface water flowoff resulting not only from the spring thawing of the snow cover, but also from the summer shower precipitation. By decreasing the flowoff, the forestation strips promote better dampening of the soil, higher levels of subsoil waters, improving thereby the moisture supply for vegetation. In order to control the flowoff, the forest strips are to be disposed athwart the slopes.

The meteorological effectiveness of the forest strips is tied in directly with their effect upon the wind velocity and, together with that, upon the turbulent exchange.

The forest strips retard the blowing off of the snow cover and promote its more uniform build-up. In the shielded fields, the snow accumulates in greater quantities than in the open steppe, which accounts for the greater moisture content in the soil during the spring thawing. The diminution in wind velocity decreases the number of sand storms and weakens their effect. In the presence of strong dry winds, the forest strips are instrumental in reducing evaporation at the expense of the greater moisture content of the soil and the more abundant vegetation, regardless of the fact that the sum total of the evaporated moisture from the forest strips and the inter-forest parcels is greater than in the open steppe.

Over the forest strips, particularly at the windward slopes,

there occurs the condensation of the so-called horizontal precipitation: hoarfrost and sleet. Under conditions of interlaced relief, this additional precipitation, which is not recorded by the pluviometer, may become considerable.

The importance of the forestation strips as direct suppliers of moisture to the atmosphere, by way of transpiration, under arid conditions, cannot and should not be too great, since accelerated transpiration is possible only at the expense of the already depleted water reserves, and may result in the lowering of the subsoil water level under the foreststrips. This, specifically, was observed by G. N. Vysotskiy in a number of places and led him to the wrong conclusion about the desiccation of the soil by ligneous vegetation in general.

The amount of moisture transpired by a forest strip, is in direct relation to the sylvan species, to the degree of denseness in planting, and other features of their growth. The problem of the agro-forestation meliorator is in the proper selection of both the tree species and the agro-technical complex, under which the amount of moisture, lost to transpiration, is to conform to the climatic resources.

We now turn to a more detailed analysis of the effect of the forestation strips upon the wind and upon the turbulent exchange connected with the latter.

As far back as 1898, N. P. Adamov, a collaborator of V. V. Dokuchayev, indicated that the effect of a forest upon the velocity of the wind is propagated approximately to a distance of about 400

meters, while the strip of calm does not exceed 10 meters in width. These conclusions were corroborated by subsequent investigations.

In recent years, the most elaborate investigations on the effect of forest strips upon the velocity of the wind were conducted at the All-Union Research Agro-melioration Institute by Ya. D. Panfilov and V. A. Bodrov, and at the Ukrainian Research Institute for Agro-forest Melioration and Forest Economy by Yu. P. Byalkevich.

The wind-shielding effect of the forest strips runs in direct ratio to their heights. The taller the forest strip, the further the propagation of its effect. Thus, the effective distance of the forest strip is measured not in meters, but by the ratio of the distance from the strip to its height.

In the mechanism of the wind-shielding effect of forest strips, a definite role is played by the loss of a part of the kinetic energy of the air current to friction and to the shaking of the tree tops. In direct proximity to the forest strip, there is also the effect of circumfluency, which increased with greater density, forming a zone of wind umbra.

Of decisive importance is the change in the structure of the air current itself sustained in passing through the forest strip, which, according to M. I. Yudin, manifests itself in the disintegration of large and the formation of small rapidly fading vortices, which leads to the weakening in the turbulent exchange and, together with that, the diminution of evaporation. Academician V. R. Williams has already pointed to the effect of vorticity upon evaporation.

Small vortices generally occur in the surface air layers, and, with height they become larger in their dimensions. It is, therefore, important that specifically the lower air layers move across the forest strip, i.e., that the forest strip is blown through in its lower part. With reference to the formation of small vortices, the effect of the forest strip is similar to the effect of a series of aerodynamic grids.

The total wind-shielding effect of a forest strip, including into this concept also its effective distance, depends, firstly, on the extent, to which the air current will be weakened in passing through the strip, and, secondly, on how rapidly it will become intermixed with the upper non-abated air current, which passed over the forest strip.

When the forest strip is blown through to a small degree only, the air current coming through it, will be weakened to its minimum value, but simultaneously the total mass of air passing through the forest strip, will also be reduced to a minimum. As a result, the upper non-weakened current beyond the forest strip rapidly descends to the terrestrial surface, and the wind-shielding effect is cancelled out at a small distance from the strip.

In the presence of a very sparse and narrow forest strip, the air current that passed through it, will be weakened to a small degree only, and, therefore, regardless of the fact that the air mass, that passed through the forest strip, will be considerable, and its complete merging with the upper undisturbed current will occur at a relatively great distance from the strip, the summary effect will also be insignificant.

The index for the shielding effect of the forest strips is the wind velocity itself, particularly in view of the fact that, with the present methods of determination the concept of wind velocity includes not only the velocity of the directional current, but also the velocities of vertical axis vortices.

Figure 70 is a graphic presentation of the wind-shielding part played by forest strips of various design. Both, the effective wind-shielding distance of the forest strips, as well as the diminution in the wind velocity at various points of the shielded zone, vary considerably. The maximum summary effect is produced by a forest strip, which is like openwork on top and sparse below. The lowest effect is produced by a forest strip, which is either continuously sparse, or continuously dense.

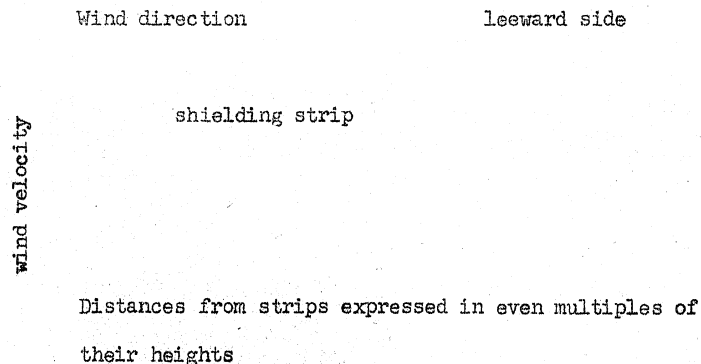


Figure 70. The effect of forest strips upon the wind velocity in percent of the wind velocity prevailing in the steppe (as per Ya. D. Panfilov). 1 - openwork below, dense on top; openwork on top, dense below; 3-dense on top and below; 4- sparse on top and below; 5-dense, 200 meters wide.

It should be pointed out that in the large gaps between the trees (in the forest strip itself), the wind velocity may be the same as in the steppe, at the expense of the diminution in the area of the transverse section of the air current.

Figure 71 is a graphic presentation of the effect of two types of forest strips upon the velocity of the wind in the 2 to 15 meter layer. With the same height of about 8 meters, strip No 10 was 6.5 meters wide at the tree tops, and consisted of 5 rows of ash-trees. The bottom three quarters was about 60 percent openwork, i.e., easily blown through. The tree tops were disposed in the upper quarter of the planting, which was openwork to the extent of 28 percent only.

Strip No 1 consisting of ash with English elm, was 49 meters wide, and the openwork characteristic along its entire height was 0 - 1 percent.

Observation were made at five levels, with the anemometers lifted to the requisite levels by a bascule of the waterwell "crane" type.

In the case of both strips, essentially extreme in their openwork characteristics, the velocity of the wind is diminished above the ridge, particularly above the wide strip. Over the narrow strip, the diminished velocities cede their place to the accelerated velocities already at the height of 1.5 to 2 meters above the tree tops. Over the wide strip, the weakening of the wind was maintained in the entire layer observed on the leeward side of the wide strip,

the sharply pronounced velocity minimum, practically down to complete calm, is to be found at the base of the forest strip edge, with the velocity rapidly increasing with distance from the latter. The narrow forest strip has two minimums: one directly beyond the tree tops at a height of 6 to 8 meters, and the other in the surface air layer at a distance equal to 6 to 7 times its height.

Strip No 10

Strip No 1

Figure 71. The effect of a sparse and dense forest strip upon the velocity of the wind (as per Yu. P. Byallovich).

1- line of equal wind velocities (velocity at the 5-meter level over the open field is taken as 100 percent); the strip of trees and undergrowth.

A. R. Konstantinov summarized the empirical data pertaining to the ratio of the summary wind-shielding effect of forest strips (at a distance equal to 30 times their height) to the degree they are being blown through. As can be seen from Figure 72, the maximum protection from the wind is afforded by strips that are blown

through to the extent of 30 percent. The best effect is produced by strips with a maximum openwork characteristic in its bottom part, which characteristic is provided by rows of mature oaks, maples, and other trees, in the absence of underbrush. Low (up to one half of a meter) creeping brushwood does not diminish the wind-shielding capacity of the strip.

Figure 72. Ratio between the total wind-shielding effect of forest strips and the degree of the openwork characteristic of the upper three quarters of the strip (as per A. R. Konstantinov).

The dispersion of the dots in Figure 72 is determined principally by the varying distribution of the openwork characteristic of the strips at the height. The upper dots pertain to forest strips with the maximum values of the openwork characteristic in the lower layer. Thus, the upper Konstantinov curve corresponds to the optimum design of the forest strips. This curve indicates, in the presence of a 30 percent-value of the openwork characteristic, a total diminution in wind velocity by 45 percent.

With reference to the optimal width of the forest strips, viewed from the angle of their wind-shielding effect, it lies, according to M. I. Yudin, within the range of 10 to 25 meters. Any further increase in width will diminish the extent to which they are blown through, diminishing thereby their wind-shielding capacity. However, the optimum width of the forest strips is determined not only by their wind-shielding effect, but also by a series of other factors, particularly by the necessity to provide for the normal growth of the trees themselves. In addition, the forest strips are to prevent erosion and washaway of the soil. Purely economic factors also play an important part. All these factors were

taken into account in the formulation of the widths recommended for the forestation strips by the Stalin plan for the transformation of the nature of our steppes.

The absolute effective distance of the strips in a leeward direction equals 30 to 40 times their heights, and in a windward direction - 10 to 12 times their heights. Thus, the effect of the forest strips is manifested at a distance equal to 40 to 50 times their heights, i. e., with the height of the trees being 10 meters, their effect will be manifested in an area 400 to 500 meters wide. The distance of the more or less effective influence (with wind velocity diminished by no less than 10 percent) extends for a distance equal to 25 to 30 times their height in a leeward direction and 0 to 3 times their height in a windward direction, altogether, then, for a distance equal to 30 times their height.

The effect of wind velocity upon the wind-shielding action of the forest strips, is manifested in different ways, in relation to the degree, to which they are blown through. For strips with a low blowing-through value, the effectiveness rises with wind velocity. For strips with a high blowing-through value, the effectiveness decreases with wind velocity (a strong wind, in bending the branches, increases the blowing-through value).

With greater turbulent exchange, the effectiveness of the forest strips is diminished, with lesser turbulent exchange - it is increased. Therefore, the minimum effectiveness of the forest strips is observed during the day, as is confirmed by the observations of Yu. P. Byallovich.

Of practical interest, is the ratio of the wind-shielding effect of the forest strips to the angle of incidence, at which the wind strikes the strip. Observations show that the departure of the wind from a direction perpendicular to the strip, within the range of 25 to 30 degrees, has no substantial effect upon its wind-shielding capacity. A forest strip, presenting to the wind a vertical wall of great roughness, weakens even a wind that blows parallel to it.

Data, collected by N. M. Gorshenin, analyzing the wind-shielding effect of a forest strip under various wind directions, is cited in Table 107. With a decrease in the angle of incidence, formed by the direction of the wind and the strip, the wind-shielding effectiveness is diminished. Yet, in the presence of a mean value of the angle of incidence equal to 11 degrees, i. e., when the wind is almost parallel to the strip, the total value for the weakening of the wind is almost 50 percent of the weakening occurring in the case of a perpendicular wind.

[see Table 107 on following page]

The wind-shielding effect of the cell formed by the forest strips in the shape of a rectangle, 400 meters wide and 1200 meters long, with the height of the strips being 15 meters, and their blow-through value being 30 percent (by the Konstantinov computation), is represented by the isolines in Figure 72a. The corresponding computation indicates that a change in the wind

TABLE 107

The weakening of the wind velocity by the forest strip in relation to wind direction.

Angle of Incidence of the wind with relation to the strip	Its mean value	Velocity of the wind (in percent of velocity in the center of the field between the strips) in relation to distance from strip, in meters.										Mean wind-shielding effectiveness in percent	
		10	30	100	200	425	200	100	30	10		Up to 200 meters	up to 425 m
68 - 90°	79°	11	27	66	92	100	104	98	89	80		42	22
45 - 68	56	12	30	71	93	100	102	98	91	85	37		19
23 - 45	34	14	45	81	96	100	105	93	93	87	30		15
8 - 23	11	30	61	89	95	100	99	94	94	86	22		12

direction by 90 degrees (from a wind perpendicular to the basic strip to a wind that is parallel to it), will cause the wind-shielding effect for the cell, as a whole, to diminish by only 10 percent. It being the case that the optimum effectiveness is observed in the presence of a wind not perpendicular to the basic forest strip, but at some angle to it.

Figure 72a. Diagrammatic representation of the wind velocity iso-lines within a solitary cell, in the presence of perpendicular and parallel directions of the wind.

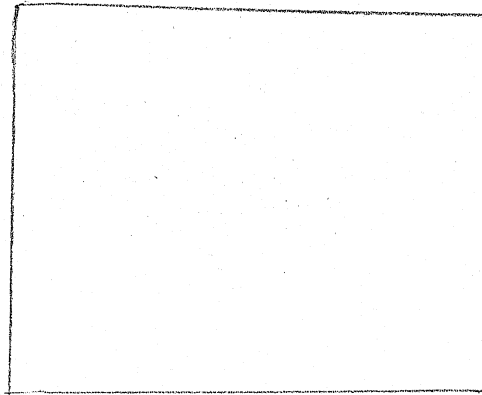
H = 15 meters; q = 30 percent; Z₀ = 20 centimeters.

One of the direct results of a diminution in wind velocity is the weakening of the pernicious effect of the "black storms", frequently observed in the Ukraine, on the Don, and in Kuban'. Strong winds, under corresponding conditions, such as in the case of large plowed up fields, which, as yet, are not covered with vegetation, force the fine particles from the soil, together with the humus, and carry them for considerable distances.

The reduction in the wind velocity brought about by the forest strips, naturally, reduce this detrimental blowing out of the soil, since the lifting power of the wind is indirect ratio to the square of its velocity. Thus, when the wind velocity is diminished from 10 to 7 meters/second, i.e., by 30 percent, its lifting power will diminish by 51 percent.

The effect of the forest strips upon the formation of the snow cover is common knowledge. By reducing wind velocities, the forest strips hinder the lifting off of the snow from the fields. A deeper snow cover on the inter-forest fields makes for more favorable conditions of hibernation of the winter crops, increases the winter moisture reserves and promotes its better absorption, since the soil, which is frozen to a lesser degree under the deeper snow cover, thaws out more rapidly and absorbs the thaw waters more completely. The effect of the forest strips upon the accumulation of snow is in direct ratio to their design. Figure 73 indicates the depth of the snow cover at various distances from the edges of 3 forest strips varying in design. According to L. A. Golubeva, the most uniform formation of the snow cover is observed at the strip that can be blown through by the wind. The dense forest

strips, and also the strips with dense undergrowth, by collecting within themselves and on the edges of large snow drifts, may not only retard the spring activities in the fields near the forest edges, but may also sustain broken trees and bushes, due to the weight of the snow.



Distance from forest strips

Figure 73. The effect on snow accumulation exerted by 8-row forest strips of various design (as per L. A. Golubeva).

1 - at a dense forest strip; 2- at a strip having an openwork pattern; 3 - at a strip that can be blown through by the wind.

An increase in the depth of the snow cover, i. e. a greater supply of winter moisture, naturally increases the moisture content of the soil, particularly at the beginning of the vegetation period. Thus, according to observations in the Rostashev district in 1936, made near a forest strip that is blown through by the wind, up to the spring thawing (March), the moisture content of the soil in the 0 to 150- centimeter layer was practically the same at

the distances of 50 meters and 500 meters from the strip, amounting to 20.5 and 19.7 percent, respectively. After the thawing out of the snow cover (the end of April), the moisture content of the soil at 50 meters from the strip was higher than 30.5 and at 500 meters from the strip only 21.4 percent. With the growth of the crops, this difference is gradually being erased, attaining full equalization only toward the end of the summer.

The wind, in controlling turbulent exchange, plays an important part in the transfer of heat and water vapor. The elimination of water vapor from the evaporating surface promotes evaporation-transpiration. It is, therefore, natural that the diminution in the wind velocity and turbulent exchange in the inter-forest areas must be reflected in the temperature, air humidity and evaporability. The diurnal temperature range of an area between forest strips, blown through only to a small degree, must become increased at the expense of an increase in the temperature maximum and a decrease in the minimum value. The absolute humidity during daylight hours has to rise, and, in connection, there-with, relative humidity may remain unchanged, and not reduced, regardless of the rise in temperature. At night, the decisive effect upon humidity may be exerted by a drop in temperature. At any rate, nocturnal relative humidity must be higher than in the open steppe.

But, in addition to the wind, the character of the vegetation itself has an effect upon the temperature and the humidity of the air. The potential possibilities for transpiration, that are present in the grass cover of the inter-forest areas, must be greater than in the open steppe with a sparser grass

cover and with a lesser reserve of moisture in the soil. This circumstance may, in its turn, raise somewhat the daylight absolute humidity, with a simultaneous drop in temperature (due to the extensive loss of heat for transpiration).

Direct air temperature observations indicate that, during the day, the effect of both these factors is, to a considerable extent, mutually compensative, and the air temperature differential between the open steppe and the inter-forest areas is not great. But in some cases, the weakening of turbulence will still be the decisive factor, and the day temperature in the inter-forest areas is considerably higher than in the open steppe, as can be seen from Table 108 (as per Bodrov).

TABLE 108

Temperature of the air over the open steppe and between the forest strips as observed in the Rostashev district
18/VIII 1934

Period of observations	Steppe	Distance from forest strip (in meters)					
		10	50	100	200	500	870
Mean for 1100,1300,1500 hours	33.8	33.8	34.8	34.8	35.0	34.0	33.9
Differential		0.0	+1.0	+1.0	+1.2	+0.2	+0.1
Sunset at 1930 hours	24.2	21.0	21.5	23.0	23.0	23.3	24.0
Differential		-3.2	-2.7	-1.2	-1.2	-0.9	-0.2

Proceeding from this, we must conclude that the forest strips in themselves do not protect the crops from the effect of the harmful high temperatures, particularly from the so-called "ignition" effect, which occurs at temperatures in excess of 35 degrees Centigrade. It must be remembered that in the case of crops with greater heat affinity, such as sunflower, maize, and others, the somewhat higher level of day temperatures may have a positive value.

With the sunset, the situation changes, and the temperature of the air between the inter-forest strips is lower than in the open steppe, which results in a somewhat accelerated frost hazard for the inter-forest strip areas (Table 108). This negative effect is linked directly to the radiation chilling and is pronounced to a full extent only in clear and calm weather (radiational freezeovers). In the presence of advection freezeovers the liability of the crops to destruction is, of course, greater in the open steppe, since the accelerated turbulent exchange promotes more rapid cooling.

The forest strips, particularly those of greater density and with undergrowth, in extending across the valley, hinder the cold air flowoff. Therefore, in the parts of the valley, disposed above the forest strip, the frost hazard is increased. In the lower parts, protected by the forest strip from the cold air flowoff along the valley, the fire hazard, on the contrary, is diminished.

The greater openwork characteristic of the lower part of the forest strips, in facilitating the flowoff of the cold air,

To analyze the duration of the unseasonal freezeovers, an additional setup of thermographs is required. Thermographs cannot replace minimum thermometers, since they are relative value devices. Thermographs are not to be set up in a Selyaninov booth, since their readings would give a distorted conception of the diurnal temperature variation. Thermographs are to be used in observation booths of the conventional type.

Figure 79. Simplified thermometer protection (designed by Borisov).

In the presence of a thermograph, the minimum and the standard thermometers are set up in the same booth on a special wire base.

The time of observation at the thermometric points, as distinct from the meteorological stations, is not established with precision, since the reading of the standard thermometer during observations is used only as a check of the minimum thermometer and thermograph, having no value in itself. This permits one observer to manage several observation points.

There may be one-time and two-time observations.

One-time observations are made about midday, from 1100 to 1500 hours, bearing in mind that the diurnal minimum temperature does not occur at this time, and, therefore, the observer will record the minimum value for the preceding night and will reset the thermometer (will move the pin to the meniscus of the alcohol), making it ready for registering the minimum temperature

partially paralyzes the frost-biting effect of the weakened turbulence. The more abrupt temperature drop at the inter-forest areas occurring in the presence of radiational chilling, is a favorable moment for days with dry wind, since, together with the drop in temperature, relative humidity is increased, weakening thereby the pernicious effect of the dry winds.

As already indicated, the sum total of evaporating moisture between the forest strips is greater than in the open steppe, but the transpiration intensity is lower, particularly on days with considerable wind velocities.

Interesting comparative data on the transpiration intensity, as between the open steppe and between the forest strips, were collected by Z. F. Samokhina at the Kamennostepnaya Experimental Station (Table 109)

TABLE 109

Amount of water lost by summer wheat in grams per one gram of absolutely dry substance (as per Z. F. Samokhina).

Observation locality	Computation days			
	27/V	12/VI	30/VI	13/VII
Open steppe.	15.44	9.48	2.84	2.92
Inter-forest field	12.60	6.44	2.16	2.16

As a result, the coefficient of transpiration (which

is the amount of water in grams lost by a plant to the formation of one gram of dry substance), under the effect of the forest strips, is decreased, while the transpiration yield (which is the amount of dry substance in grams formed per one kilogram of water consumed) is correspondingly increased. This is confirmed by the experiment of growing summer wheat in containers with soil of the same moisture content.

Table 110 shows the results of this experiment (by G. I. Matyakin), which results are integral exponents of the "summer" effect of the forest strips upon the complex of meteorological conditions, which directly affect the yield of agricultural crops.

TABLE 110

Transpiration coefficient and yield computed by
data obtained through vegetation experiments

Distance from forest strip (in meters)	Transpiration coefficient	Transpiration yield
10 - 15	482	2.08
30 - 35	455	2.20
100 - 105	483	2.09
150 - 155	660	1.52
400 - 405	622	1.61

Further research and experimentation will result in still

greater effectiveness of the forestation strips as applied to various conditions of climate, terrestrial relief, and soil. It will also result in a detailed quantitative analysis of their effect upon the elements of climate and weather.

The microclimate of the forestation strips themselves approximates the microclimate of a forest, but can vary substantially, in relation to its structural design, its denseness, and the presence of undergrowth. Just a few words relating to the microclimate of the forestation strips in the first years of their growth, in connection with the "cradle" method of their planting proposed by academician T. D. Lyenko. The "cradle" planting of oak, 35 to 40 acorns to an area of one square meter, in a row with maize or sunflower, and between the rows of any plowed-through or grain crop, provides for the young oaks a favorable microclimate, primarily some shade and protection from wind. During the winter, the stems of these plants are instrumental in snow detaining, thus providing moisture for the second year of growth not only for the young oaks, but also for the newly planted, together with the winter rye, maple and undergrowth.

In planting forestation strips by the method of T. D. Lyenko, the young trees, during the first four years of growth, when they are particularly exacting with relation to their environment, find themselves under the protection of cultivated herbaceous vegetation, which is instrumental in creating a favorable microclimate for their successful growth.

DIVISION VI

METHODS OF RESEARCH INTO LOCAL CLIMATE AND MICROCLIMATE

CHAPTER 20

RESEARCH PROCEDURES

To date almost every researcher in the field of local climate and microclimate was beginning his work by trying to resolve questions relating to methods as applicable to the problems facing him, frequently considering it unnecessary to bring them into harmony with the methods and procedures of others.

However, with the development of science, the attainments of the latter became accessible to ever wider technical utilization, which, naturally, is accompanied by a standardization of methods. Such standardization is also indicated for the methods of studying local climate and microclimate.

Below are cited the methods, which were used by the former Agrohydro-meteorological Institute VASKhNIL, by the Climatology Division of the Geographical Faculty of the Leningrad State University, and by the Central Geophysical Observatory.

Before proceeding with the analysis of concrete methods, let us turn to the general problems of the organization of the studies of microclimate and local climate, considering that their success depends not only on the perfection of apparatus and the precision of calculations, but primarily on the proper procedure of the entire investigation and its purpose.

In arranging the studies, it is necessary to formulate clearly the appointed assignment. The complex of meteorological phenomena is very involved and multiform, as also is the interaction between the climate and other physico-geographical factors.

This makes the concretization of the assignment, the separating out of the basic object of the study very difficult, but at the same time absolutely necessary, in order to arrive at conclusions in a minimum period of time.

The end result of each study - a practical utilization of the conclusions arrived at, runs in direct ratio to the fixed time of its execution.

It follows therefrom, that even only an approximated, yet well-timed solution of a problem has absolute preeminence over a precise solution arriving too late.

Which, then, is the guide line to be followed in the concretization of an assignment?

It is necessary to differentiate between two varieties of studies:

(1) meteorological phenomena are studied not per se, but as an environment, on which other phenomena are dependent;

(2) a specific meteorological phenomenon is considered as the basic object of the study, and its relation to physico-geographical conditions is exposed.

The second variety of study is essentially a part of the first one. However, in individual most difficult cases, or when the

phenomenon involved is of practical interest to a number of branches of the national economy, it becomes rational to segregate the study of this phenomenon into an independent assignment.

As an example, such a phenomenon as the foehn can be cited. The foehn is of practical interest to the agricultural, forest, and health resort economies, and, in addition it exerts an effect upon the thawing of the snow cover and glaciers. It is, therefore, more effective to first study the general regularities of this phenomenon in relation to the weather and to the physico-geographical conditions only, without the direct calculation of its effect upon any of the branches of the national economy, but with the ever present object to apply the conclusions arrived at to practical industrial assignments.

In attempting the solution of the first problem, it is first of all necessary to ascertain, which meteorological elements, during which seasons of the year, and which hours of the diurnal period, at what level, in the air or in the depth of the soil, exert a decisive effect upon the object, which is under our study.

Let us cite an example. The propagation of subtropical crops is limited by the frost hazard, it being the case that the main danger lies not in the spring-autumn freezeovers, which can only damage the crop, but in the winter frosts, under the effect of which subtropical crops, such as lemons, mandarins, tea, may perish completely, unless proper protective measures are taken.

It is, therefore, natural that the study of the frost hazard of a territory is to be timed to the winter season.

In addition, the basic observations must be timed to the nocturnal hours, since, in the subtropics, due to a low latitude a well pronounced diurnal temperature variation is manifested even during the winter.

The surface layer temperature inversion, forming at night, is dissipated during the day, and, in place of the big differential of the low nocturnal temperatures, a relatively uniform thermal cycle is established during the day.

Because of this circumstance, daylight observations even on days with frost cannot uncover all the diversity in the degree of the frost hazard, and therefore, regardless of the technical difficulties, the basic meteorological observations in this case must be conducted at night, including those nights, when the temperature, although above zero, yet retains the general character of a distribution peculiar to the critical frosts during clear and calm weather.

Having ascertained the specific meteorological elements to be studied, it becomes necessary to formulate a working theory on the regularities of the phenomenon under study, with the full utilization of the available experience and the general theoretical concepts. In the contemplated example it was necessary to assume the character of the change in the minimum temperatures in relation

to local conditions. On the basis of the existing analysis by meteorological stations, it was supposed that in Transcaucasia the distribution of minimum temperatures is determined by terrestrial relief, distance from the sea and altitude. In addition, it was accepted as a working theory that the effect of terrestrial relief upon the minimum temperatures is basically controlled by its effect upon the flow off and the stagnancy of the cold air during the period of nocturnal inversions (i.e., by its convex and concave form), and that the slope exposure (southern and northern slopes) has no direct effect upon the distribution of the minimum temperatures. In addition, it was established preliminarily that the moderating effect of the sea is diminished with particular sharpness in direct proximity to the latter.

The assignment consisted in checking the above suppositions and determining their quantitative expressions.

It follows from the above that, before embarking directly upon the field studies of microclimate or local climate, it is necessary to perform extensive preliminary work, the successful completion of which spells out the success of the complete assignment to the extent of 50 percent.

The first stage of the study calls for a general familiarization with the territory and its physico-geographical features by way of a few observations, conducted by the researcher personally, by way of questioning local functionaries, who, in the performance

of their daily work, come into contact with nature. This preliminary field reconnaissance helps in the final formulation of the assignment, in the selection of the place of observation, and in the more precise specification of the volume and methods of observations. All this creates the background for pursuing the study successfully in its second and basic phase.

A climatologist studies not an isolated physical phenomenon in the atmosphere, but the interdependent relationship of this particular phenomenon with all the other elements in the landscape. He, therefore, conducts, alongside of meteorological observations, observations of the physico-geographical and biological phenomena, which are linked to the above. Geographic coordinates, altitude, the form of terrestrial relief, the soil, and the vegetation must be described, and their dynamics taken into consideration.

The description of the indicated landscape elements is done from the angle of their effect upon the microclimate and local climate element under study, and, in the case of an assignment of the first type, also from the angle of the possible climatic effect upon them.

The study of the physico-geographical complex is to be concluded by its classification by the climatic criterion. Without such a classification, the results of field climatic studies will have a limited local significance only, and the conclusions drawn will not be adequately applicable to practical utilization.

It is quite impossible to physically embrace all the gardens, all the valleys, all the oases through direct observations. Instead, we investigate typical localities, and then extrapolate our conclusions upon analogous conditions elsewhere, using the above-mentioned classification for the establishment of said analogies.

Chapter 21

PRECISION AND REPRESENTATIVENESS OF OBSERVATIONS

The problem of precision in observations is lined directly with the problem of their representativeness.

It is necessary to differentiate between the following: the precision of the instrument reading, the precision of individual observations, the precision of the mean values, the representativeness of individual observations both in time and in space, the representativeness of averaged data both in time and in space.

The precision of the instrument reading is usually determined by its scale. Only in the presence of a rapid change of an element in time, the latter affects the precision of the count. As an example, the effect on the precision of the count of the oscillations of the needle of a galvanometer, which is connected to a non-reluctance resistance thermometer. The oscillations of the needle in this case can be so wide and continuous that the eye will not grasp its position in a definite moment of time.

The precision of individual observations does not, by far, coincide with the precision of the count, since all instruments have inherent errors, systematic and casual. Alongside of the inherent errors in the instrument, due to the inadequate precision in its fabrication or calibration, it is necessary to take into account the errors attributable to the peculiarities of its design,

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its installation, and the method of its utilization.

As an example, we point to the conventional mercury thermometer. As any thermometer in general, the mercury thermometer registers its own temperature only. In order that it indicates the temperature of some other body, it is necessary that it assumes precisely the temperature of that body, which condition is not always easily attainable. In order that it indicates the temperature of the air, it has to be shielded from the radiational heat exchange by being placed into a special booth or a ventilated instrument case (ventilation psychrometer). The temperature of the air in the booth is different from the temperature of the outside air; during the day it is higher, and at night it is lower. The readings of an exposed thermometer differ even to a greater extent from the air temperature. The so-called radiational error of the thermometer may attain the value of 3 degrees Centigrade and even more. Therefore, the precision of the observation of the temperature of the air is usually considerably below the precision of the count.

It is necessary to differentiate between the casual and the systematic errors. The value of a casual error, which is determinable by casual departures of an individual observation from the true value of the element being measured, and is linked to the design characteristic of the instrument, as a rule, is not in excess of certain limits, peculiar to the given instrument or to the given method. The narrower the variation limits of the casual errors, the higher the quality of the instrument. The averaging

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of the data, in the presence of a casual error, increases the precision of the observation results.

The reduced formula of the mean arithmetical error m in the case, when the departures are casual, has the following aspect:

$$m = \frac{\sigma}{\sqrt{n}}, \quad (27)$$

where σ is the mean quadratic departure; n is the number of observations.

The case of a systematic error is entirely different. It is also linked to the design peculiarities of the instrument, and it occurs in two varieties.

Usually, a systematic error is distinguished by its stability. The averaging of observations with a stable error does not reduce its distorting effect.

The second variety is constituted by errors, which change their value with regularity. The radiational error of thermometer readings belongs to the classification of varying systematic errors. Its daylight value is usually positive, its nocturnal value -- negative. It is greater under a clear sky and smaller under an overcast sky. The averaging of this error may be useful only in individual cases. For example, in computing the mean diurnal temperature, the positive error of the daylight observations

is partly cancelled out by the negative error of the nocturnal observations. However, in determining the diurnal temperature range by the temperature maximum and minimum, the averaging of the observations does not, to any degree, reduce the radiational error.

In the case, when we need not the absolute magnitudes, but their differences (derivatives, differentials), the stable systematic error is leveled off. In computing the differences, it is automatically eliminated:

$$(A + m) - (B + m) = A - B, \quad (28)$$

where m is the stable systematic error.

Such is not the case when we deal with ratios of magnitudes, as, for example, with the ratio of wind velocities when analyzing their variations with height, as

$$\frac{A - m}{B + m} \neq \frac{A}{B} \quad (29)$$

In this case, the distorting effect of the error will be the greater, the greater the extent, to which the ratio $\frac{A}{B}$ differs from unity.

When analyzing by the method of ratios, it is best that the relative value of the error remains, i.e.:

$$\frac{m_1}{m_2} = \frac{A}{B},$$

where m_1 is the error for A, m_2 -- the error for B, since in this case

$$\frac{A + m_1}{B + m_2} = \frac{A}{B}. \quad (30)$$

As an example of retaining the uniform relative error (expressed in percent of the absolute value of the magnitude being measured), we can cite the errors tied in with the reduction factors in the conversion from the number of graduations of an instrument, let us say, a galvanometer, to absolute magnitudes, such as calories. The absolute error in this case grows with the increase in the number of graduations, while the relative error remains unchanged.

The stability of systematic errors is of great importance in all those cases, when the determination of their value and the interpolation of corresponding corrections are for some reason impossible.

Practically, the preservation of the stability of corrections may be accomplished by the utilization, for the entire series of observations, of the same instrument, or, at least, instruments of the same design, and also by maintaining, to the maximum possible extent, the homogeneity of the accompanying phenomena.

It is specifically due to such considerations that no determination of the vertical temperature gradient by using, at one altitude, a thermometer installed in a psychrometric booth, and, at another altitude, a ventilation psychrometer, is permissible.

In using instruments of various design, we are dealing not only with inherent errors directly, but also with such characteristics of instruments, due to which observations, even though precise unto themselves, turn out to be non-comparable between each other. A great part, in this respect, is played by the reluctance characteristic of the instrument and the scale of averaging tied in with the latter.

The thermometer in the psychrometric booth has more reluctance than the thermometer in the ventilation psychrometer. As a result, individual counts by both instruments, even with the interpolation of the correction for the radiational error, will be non-comparable, since they will represent temperature values averaged over different time intervals. The same picture is observed in the parallel use of a psychrometer and a non-reluctance platinum resistance thermometer.

The parallel variation of readings by the two thermometers, presented in Figure 30 (page 68 of original), visually illustrates not only the different range of the oscillations, but also the relative lag in the readings of the ventilation psychrometer. But even instruments of the same design may register various

systematic errors. This is the reason for the use of the same instrument, if at all possible, and for shifting instruments about, when it becomes necessary.

Instruments are shifted in the observations of the variation of wind velocity with height. For example, in the case of a more precise determination of the ratios of wind velocities at the 1- and 2-meter level ($\frac{U_2}{U_1} = R$), two series of observations following each other are made, it being the case that, after the first series of observations, the instruments are shifted around. The velocities for each level are obtained by averaging the count of the readings of both series, in which case the errors in the velocity readings being compared are automatically averaged and smoothened. It is, however true, that, in ratios of magnitudes, equal errors are not fully eliminated. R. E. Soloveychik, proceeding from the theory of probability, maintained that such shifting of instruments make for a higher degree of precision in all cases, when the ratios of velocities (or other contemplated elements) lie within the limits from 1/3 to 3.

All errors, regardless of their nature, may be presented in absolute and relative terms. An absolute error is expressed in the same units as the contemplated magnitude, while a relative error is computed as a ratio of the absolute error to the contemplated magnitude, and is usually expressed in percent. In some cases, it is more rational to compute the value of the relative error in ratios not to the absolute magnitude of the contemplated

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element, but to the magnitude of the possible variation of the latter. For instance, we measure pressure with a precision to 0.1 millibar, which results in a relative error of approximately 0.01 percent. But on the terrestrial surface, pressure varies approximately within the range of 100 millibars. Therefore, in relation to the possible variation of pressure, the above relative error will be 10 times greater, equaling 0.1 percent.

In selecting the method of research, it is useful to remember that the relative error δ of a sum is equal to the ratio of the sum of the absolute errors Δ to the sum of the contemplated magnitudes:

$$\delta(x+y) = \frac{\Delta x + \Delta y}{x + y}, \quad (31)$$

but the relative error of the difference equals the ratio of the sum of the absolute errors to the difference of the contemplated magnitudes:

$$\delta(x-y) = \frac{\Delta x + \Delta y}{x - y}. \quad (32)$$

Therefore, in the presence of even relatively small errors in the minuend and the subtrahend, the relative error of the difference may be considerable.

For example, it is disadvantageous to determine the effective emission from the terrestrial surface as the difference between the emission of the earth itself and the back radiation of the firmament. Both these magnitudes are rather considerable, but the difference between them is small. As a result, the relative error of the difference, which in this case is the value of effective emission, may attain 100 percent or even more (if an error of 5 to 10 percent has been made in the determination of the emission from the earth and of the back radiation from the sky).

Precision in measuring does not yet guarantee the typicalness of the obtained value for the contemplated phenomenon. The measured value must be representative (representative in precise translation means typical, significative, characteristic).

In determining representativeness, it is necessary to proceed from the assigned purpose of the observations. If the local climate of a moss swamp is the object of a study, the area of observation must be typical for a swamp not only by the characteristics of the active surface, but also by the structure of the adjacent air layers, which structure may be subject to deformation by surrounding objects, such as a nearby railroad embankment or trees.

A good index of the structure of the surface air layers is the vertical configuration of the wind. In an exposed level locality, in windy and overcast weather, the wind velocity variation

with height follows a logarithmic regularity. The departure of the wind configuration from the logarithmic is usually tied in with local deceleration or acceleration of the wind and, consequently, the turbulent exchange, which affects the thermal and moisture exchange between the soil and the air, and together with that, the entire microclimate and local climate.

Representativeness is also distorted by local advection, for instance, by the downflow of cold air from an adjacent slope, only in the case, of course, when the indicated advection does not constitute a typical feature in itself, as, for example, when the local climate of synclines is under study.

To the category of non-representative observations can be relegated the observations of some so-called "forest stations" located not within the forest itself, but in glades of various dimensions. Such observations characterize the local climate of the glade, which, as is known, differs from the climate of the forest, with relation to some climatic elements (as the diurnal temperature variation), to a greater degree than the exposed steppe areas.

Also non-representative for cultivated fields under crops are temperature observations made on soil under its natural surface.

However, the correct selection of the observation locality does not yet guarantee the representativeness of individual observations, which may be distorted by microoscillations in both time and space.

The above cited Figure 30 in the previous text depicts the value of the air temperature microoscillations in time.

For the analysis of microoscillations in space, data, collected by N. N. Trankovich and shown in Table 111, can be used.

All the 16 thermometers are installed on the same terrace, a distance of one meter between them. Nevertheless, the readings, as between the outer ones, show a difference of 5 degrees Centigrade. The temperature oscillations attain even greater values on the actual surface of the soil.

[See next page for Table 111]

Each one of the tabulated readings is correct, but, by far, not all of them are representative. ~~It being the case that,~~ In order to make it representative, several readings in each case are required.

The variability of the meteorological elements in time, and the implicit non-representativeness of individual instrument readings, which is attributable to it, are long since being taken into account by meteorologists. In the "Instructions to meteorological stations and outposts" it is specifically underscored that each series of actinometric observations is to incorporate from 3 to 5 readings of the instrument [per individual observation]. However, the necessity to also take into account the variability

TABLE 111

The readings of 16 thermometers at the 5-centimeter
depth in the soil at 1200 hours 28-29 May 1935 taken
at the Far East Experimental Station AGMI

	No of thermometers																Mean value
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	
Temperature	21.5	20.1	18.7	20.5	20.6	20.1	16.5	18.5	20.1	20.5	20.1	19.8	21.3	20.6	18.3	20.1	19.8
Departure from mean	1.7	0.3	-1.1	0.7	0.8	0.3	-3.3	-1.3	0.3	0.7	0.3	0.0	1.5	0.8	-1.5	0.3	±0.9

of the meteorological elements in space has not as yet penetrated into the procedures of meteorological research, notwithstanding the fact that in some cases, such as temperature observations in the upper layers of the soil, it is indubitably called for.

Chapter 22

THE METHODS FOR MICROCLIMATIC OBSERVATIONS -
INSTRUMENTS

In selecting and evaluating instruments for microclimatic observations, it is necessary to be guided by the well-known characteristics of the lowest air layer, namely: the exceptionally large vertical gradients, high horizontal variability, micro-oscillations in time, low wind velocities, and decelerated turbulence. In connection with the above, specifications for higher precision (as compared to instruments for the observation of general meteorological phenomena) are not always required.

We shall not engaged in the description of the instruments, since it has already been done in the respective textbooks. But we will evaluate some of them from the angle of their suitability for microclimatic observations, paying particular attention to instruments that are already in wide use.

The leading elements of microclimate are the temperature of the lowest air layer and the upper layers of the soil, on which elements we shall concentrate our attention. Anemometers will be discussed in the chapter devoted to anemometric survey.

Specifications to be satisfied by thermometric apparatus for the lowest air layers may be formulated as follows:

(1) The instrument (or the perceiving part of it, if it can be carried out to a distance of several meters from the remaining part of the instrument), is to be sufficiently small so that the processes studied are not distorted by its very presence.

(2) The instrument is to be as free as possible from radiational error.

(3) In the special observations for microoscillations, the instrument is to have a minimum reluctance value, which feature, however, is not desirable in all other cases, since it complicates observations.

(4) To avoid the disturbance of the vegetation cover during observations, the instruments are to be of the remote-control variety.

Preference is to be given to self-recording instruments.

The above enumerated specifications are well satisfied by electric instruments, such as resistance thermometers and thermometers with thermocouples (for the descriptions and diagrams of which see V. N. Kerolivanskiy, Meteorological Instruments, 1947). Of particular importance are the features of remote control and of self-recording. In addition, they are irreplaceable for the study of microoscillations.

It must be underscored that electric thermometers, capable of recording microoscillations, usually have an insignificant radiational error. For the purpose of bringing out the mean

characteristics, the perceiving part of the instrument is to be made more massive, which causes an abrupt rise in the radiational error. This factor, and also a number of technical difficulties inherent in the utilization of this instrument under field conditions, explains the fact why electric thermometers, up to recently, were not used by us for the systematic observations of air temperatures, but only for methodological research, particularly for the evaluation of the degree of precision obtainable with ventilation psychrometers. The latter, at this time, are the only instruments suitable for wide use.

The double shielding of the thermometer wells with luminous nickel-plated cylinders, in the presence of ventilation, considerably reduces the radiational error of the ventilation psychrometer. However, the ventilation itself is a source of new errors, attributable to the considerable layer of air being sucked into the device.

Let us dwell in some more detail on these errors, inasmuch as they are of a general methodological interest. The radiational error is a result of the heating of the instrument by the sun during the day and the radiational chilling of the same during the night. Since the radiational error is accelerated with a diminution in turbulence, its value will be particularly high at the terrestrial surface, in the midst of the grass cover, in clear and calm weather.

In the process of ventilation, a rather considerable volume of air, regardless of whether the instrument is placed horizontally

or vertically, is drawn through the psychrometer, as a result of which the thermometer will record the mean temperature of some layer of air, measured in tens of centimeters. This means that, at the very terrestrial surface, the readings of the psychrometer will pertain to a greater altitude than the one, on which the device is mounted. The ventilation error increases with proximity to the surface and with the increase in the value of the gradient. It follows that, although the presence of ventilation almost completely eliminates the radiational error of the thermometers, the data recorded by the ventilation psychrometer will distort the concept of the gradients in the very lowest air layer (10-20 centimeters) by underestimating them.

In keeping with its design, the ventilation psychrometer should be suspended vertically for purposes of observation. However, in its vertical position at small altitudes above the soil, the reading of its recordings, is difficult. It is, therefore, for purposes of microclimatic observations, placed in a horizontal position. Comparative observations have registered no differences in the readings, as between the horizontal and the vertical positions of the psychrometers, and the problem is reduced to the necessity of maintaining the normal functioning of the ventilation device. If the ventilation velocity in the horizontal position is maintained normal, there is no objection to such a position for purposes of observation. It is only necessary to forestall the solar rays from penetrating inside the cylindrical shield.

The ventilation psychrometer, to date, is the only suitable instrument for observations of the humidity of the lowest air layers.

For air temperature and humidity observations above the layer of 1.5 - 2 meters, a tilting ventilation psychrometer, of the type designed by N. N. Zubov in 1923, is used. In the tilting psychrometer, in place of the conventional thermometers, are inserted thermometers with a disconnecting mercury column, of a type similar to the deep-water hydrological thermometers, with scale graduations from the end of a capillary, opposite the bulb. The psychrometer itself is placed in ^{ed} a special frame, in which it can be, when so desirable, inverted. The Central Geophysical Observatory conducted successful tests with tilting psychrometers at altitudes up to 15 meters.

Mercury and alcohol thermometers, when shaded by various shields, but without ventilation, are the least precise devices for the determination of the temperature of the air.

[See next page for Table 112]

As can be seen from Table 112 (as per E. V. Kontsevich), the non-shielded thermometer records a higher temperature than the psychrometer during the day and part of the morning, while in the evening (after the sunset), it records a lower temperature, with the differences oscillating, in some cases, attaining the value of 3 degrees Centigrade.

TABLE 112

Temperature differences by the readings of a ventilation
psychrometer and a non-shielded thermometer in clear (I),
semi-overcast (II) and overcast (III) weather at the height
of 1 meter -- Pushkino, 1934

Differences	0600 - 0900 hours			1100 - 1600 hours			1900 - 2200 hours		
	I	II	III	I	II	III	I	II	III
Mean	-0.9	-0.2	0.0	-1.5	-0.3	-0.4	0.6	0.1	0.3
Maximum	-2.4	-0.3	0.6	-3.0	-0.5	-1.8	0.8	0.3	0.4
Minimum	-0.4	0.0	0.1	-0.8	-0.2	-0.1	0.4	0.0	0.2
Number of observations	36	21	18	41	17	17	33	22	20

TABLE 113

Temperature differences by the readings of a ventilation psychrometer and shielded thermometers in clear weather at the height of 10 centimeters

	0600 - 0800 hours			1100 - 1600 hours			1900 - 2200 hours		
	mean	max	min	mean	max	min	mean	max	min
Double brass apron	-1.3	-1.8	-0.7	-2.8	-3.0	-2.0	0.7	1.1	0.3
Double paper	-1.8	-2.9	-0.9	-4.0	-5.0	-2.6	0.9	1.3	0.4
Ordinary paper	-1.4	-2.1	-0.6	-2.9	-4.3	-2.1	0.8	1.5	0.4
Open paper (semicone)	-1.2	-3.0	-0.7	-2.0	-3.9	-1.2	0.8	1.6	0.4
Horizontal wooden disk	-1.2	-1.9	-0.7	-2.7	-3.3	-1.8	0.5	0.8	0.2
Non-shielded thermometer	-1.2	-3.0	-0.5	-2.5	-4.1	-1.5	0.9	1.3	0.1
Number of observations	35			35			30		

At lower altitudes, the departures in the readings of non-shielded thermometers from the actual air temperatures should increase, due to the reduction in the coefficient of turbulent exchange.

As seen from Table 113 (in accordance with E. V. Kontsevich), all types of shielding during the daylight hours result in readings with greater departures from the readings of the ventilation psychrometer than the ones indicated by a non-shielded thermometer. This was to be expected, since their albedo is, of course, of a lower value than the albedo of the mirror-like surface of the non-shielded mercury thermometer bulb, not counting the fact that the immediate proximity of the shielding to the bulb is instrumental in the transfer of some additional heat to the latter.

During the nocturnal hours, the differences are diminished, and the minimum error is produced by the horizontal disk shielding, the so-called Borisov-simplified shielding. It is made up of two disks, 8 and 12 centimeters in diameter placed over the thermometer bulb (see Figure 81 in the text to follow). These disks, while preventing the direct radiational exchange between the thermometer bulb and the upper layers of the atmosphere, do not hinder the radiational exchange between the thermometer bulb and the surface of the soil. This, specifically, is the circumstance, which causes the nocturnal readings of the disk-shielded thermometer to be closer to the actual air temperature. The

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Borisov-simplified shielding may be recommended for minimum air temperature determinations in the midst of a grass cover.

Some researchers use thermographs of the generally accepted type for their microclimatic observations. From the analysis at the beginning of this chapter it should be clear that such an unwieldy device, as the thermograph, requiring a still more unwieldy shielding arrangement, can under no circumstances be used for observations in the midst of a grass cover. The thermograph creates around itself a microclimate of its own.

The basic specifications to be satisfied by apparatus for measuring the temperatures of the soil are as follows:

(1) A minimum disturbance of the soil and grass cover, in both the mounting of the instruments and in the conducting of the observations, is to be observed.

(2) A most complete contact to be maintained with the soil at the predetermined depth, while insulated from the thermal effect of the contiguous layers of the soil, the air, and the radiational processes.

(3) An adequate strength of the instrument in keeping with exceptional severity of conditions (diurnal temperatures ranges up to 40 degrees Centigrade, humidity ranges -- from an air-dry condition to complete moisture saturation), and also with the mechanical effects on the part of the soil in its distension, its cracking, its freezing, and its thawing.

(4) Simplicity in the mounting and in the care of the instrument for mass observations.

The most widespread are the soil thermometers of the S. I. Savinov type, which are of angular shape, due to which, when the upper part of the thermometer is at a 45-degree angle to the surface of the soil, its lower part together with the bulb assumes a horizontal position, and thus occupies the least vertical expansion. With such disposition of the thermometer, its readings conform to the greatest degree with the predetermined depth, and, in addition, the structure of the soil directly over the thermometer remains undisturbed. The horizontal disposition of the thermometer bulb acquires a particular importance, when it is elongated. However, in the Savinov thermometers, fabricated at the present time, the bulbs are much foreshortened -- their diameter is equal or practically equal to their length. By virtue of this, the bulb, regardless of its position in the soil, also extends practically the same vertical distance, and, consequently, one of the advantages of this design is cancelled out to a considerable extent. One of the deficiencies of this design is that its setting up unavoidably disturbs the soil and vegetation cover. This disturbance has a particularly negative effect upon the subsequent readings of the thermometers, particularly during the height or toward the end of the vegetation period.

Taking this into account, the Section of Climatology of the Sablinsk Scientific and Training Station LGU, in its research, conducted as far back as 1937, rejected the generally accepted angular-shape Savinov thermometer and replaced them with straight

thermometers, in which all the other features of the Savinov device were retained. The straight shape of the thermometer simplifies its fabrication considerably, reduces the percent of rejections in production, which, undoubtedly, is of practical importance. The change in the design was in its time coordinated with the inventor (S. I. Savinov).

In setting up the straight soil thermometers, a wooden board T-square is used to fix the angle of inclination of the thermometer. The thermometer is usually set at a 60-degree angle of inclination, and, for purposes of stability, it is supported by a light strut. A hole in the soil, down to a predetermined depth, is preliminarily forced with the aid of a rod having the same diameter as the thermometer. Since the thermometer is being set at an angle of 60 degrees to the surface of the soil, the length of rod being inserted into the soil, and also the length of the thermometer, are to equal the specified depth divided by $\sin 60^\circ = 0.87$.

Experience shows that, for shallow soil depths, standard thermometers with cylindrical bulbs, also attached thermometers, and sling thermometers (of course, without holders) can be used. All these thermometers react sharply to mechanical disturbance. The freezing and the thawing of the soil result in their disintegration, thus eliminating the time of their application to the periods of stable temperature above zero Centigrade. The low thermal conduction of glass provides for the relative insulation from the thermal effect of both the upper layers of the soil and from the sun.

For observations, under field conditions, up to soil depths of 0.5 meter, the feeler-thermometer designed by B. G. Ivanov, can be used.

The feeler-thermometer is a wooden rod into the end of which is fitted a small thermometer of the sling type. A metal cap, filled with petroleum asphalt, is fitted over the thermometer bulb. The petroleum asphalt prevents the thermometer from breaking and imparts to it a high degree of reluctance. The feeler is inserted into a vertical hole, driven by a wooden or metal rod, which has the same shape as the feeler-thermometer. When the soil is very compact, the rod is driven into the soil with the aid of a wooden mallet.

The determination of the temperature of the surface layers of the soil under the snow cover, including the node of fasciculation, is particularly difficult.

When the snow cover is of considerable depth, the approximated temperature readings are obtained with the aid of an extension thermometer.

For episodic round-the-clock temperature observations of the snow cover, conventional standard thermometers are used, by inserting them horizontally into a fresh cut made so that it faces north. The cut in the snow cover is trimmed every hour.

It is regrettable that all the above described instruments are not remote control devices. In this respect, electric ther-

monometers are of great advantage. But, similarly to the case of the electric thermometers for recording the air temperature, electric thermometers for the soil are so far available only as experimental samples, and, in most cases, are not suitable for use under field conditions. One of their important deficiencies is that occurring damages cannot be detected as readily as in mercury thermometers, and sometimes results have to be rejected at the completion of observations, due to errors accumulated by damage to the instruments, not having been detected in time.

The most widely used electrical thermometers for recording the temperature of the soil are resistance thermometers of various design. The resistance thermometer of the Tret'yakov type is a copper wire collector, the wire diameter being 0.05 millimeter, with a resistance of approximately 50 ohms, the wire unit placed into a hermetically sealed tube, 6 millimeters in diameter. The thermometer operates as per the diagram of the unbalanced bridge joint. The galvanometer departures are graduated to read directly in degrees Centigrade. With the aid of a switch, the resistance collectors located at various depths, are brought into the galvanometer circuit. The device receives its current from a 2 to 4.5 volt dry cell batteries. With the aid of a suitable rheostat control, the force of the current in the device is read by the red line in the galvanometer scale.

The remote control feature of electric instruments (all the measuring and recording paraphernalia may be indoors) makes them irreplaceable not only during the winter, but also

in the summer, when the direct readings at the observation point disturb and damage the grass cover, introducing thereby a considerable element of error into the analysis of the thermal cycle of the soil.

Temperature observations, pertaining to the surface of the soil, are of particular interest. It is specifically the surface of the soil that determines the heat exchange value soil-air. Important microorganisms are developed on the surface of the soil, and, through the surface, pass into the soil the stems of the plants. In addition, the soil surface temperature serves as an indirect index of the temperature of the surface of other bodies, which, like the soil itself, receive and emit heat not only by way of turbulent exchange, but also radiationally. It is known that the air temperature alone does not always provide the final characteristic of climate as an environment. Alongside of the temperature "in the shade" (air temperature evaluation by observations in the booth), "in the sun" temperature must be taken into account. The temperature of the surface of the soil is specifically utilized as an index of the "in the sun" air temperature.

Over a period of decades, the soil surface temperature was determined with the aid of liquid thermometers, the thermometer bulb inserted into the soil to its half-way mark. It is known that any thermometer records its own temperature only. In the case of liquid thermometers, this inconsistency is further

complicated by the fact that the thermometer bulb, being a 3-dimensional body, cannot truly reflect the temperature of the 3-dimensional surface of the soil, which attains its maximum during the daylight heating, and drops to its minimum value in the nocturnal chilling. As a result of this, the surface thermometer underestimates the day temperature and overestimates the night temperature.

There is, however, another source of errors. The temperature of the surface thermometer bulb is determined by its heat exchange value with the soil, being effected over an area equal to half of its entire surface, while the upper half of its surface is in contact with the air, and, in addition, receives and emits heat through radiational exchange. The coefficients of emission of both the glass and the surface of the soil are not essentially different from those of a black body. However, the albedo of denuded soil is approximately 15, rarely 30 percent, while the mirror-like surface of the thermometer bulb has an albedo of over 90 percent. Thus, if the error in the nocturnal readings of a surface thermometer is insignificant, the "in the sun" readings during the day are highly underestimated. When the surface of soil is damp, the loss of heat to evaporation, which is absent in the thermometer bulb, will reduce the temperature of the soil and, in some measure, compensate for the radiational error. When the surface of the soil is dry, and evaporation diminished sharply, this compensatory value is reduced to a minimum. Since high temperatures usually occur in the presence of considerable

aridity in the soil, it may be asserted that, for ultimately high temperatures, this compensatory value will be at its minimum, and, consequently, the actual soil temperature, during these moments, will be considerably in excess of the thermometer readings.

The temperature of the soil surface can also be obtained with the aid of the so-called thermal netting. The latter is a metal frame, inside which there are ordinary coils of platinum wire, 0.07 millimeter in diameter, fastened to the frame by celluloid strips, providing for proper insulation. When readings are to be taken, the frame is placed in such a way that the wire coils are laid out completely on the surface of the soil. The netting is switched in as one of the branches in the balanced bridge diagram. Parallel observations by the thermonetting and mercury thermometers made it possible for L. I. Zubenok to determine the differences between the readings of the netting and the mercury thermometer $\theta_s - \theta_r$. These differentials turned out to be in direct ratio to the differentials between the soil surface temperatures, as read by mercury thermometers, and air temperatures at the height of 150 centimeters:

$$\theta_s - \theta_r = 0.40 \theta_r - \theta_{150} - \text{for sand} \quad (33)$$

$$\theta_s - \theta_r = 0.56 \theta_r - \theta_{150} - \text{for plowed field} \quad (34)$$

where θ_s is the temperature of the soil surface by the thermonetting;

θ_r is the same by the mercury thermometer; θ_{150} is the air temperature at the height of 150 centimeters.

In the presence of usual daylight temperature differentials, by the soil surface thermometer and the air temperature reading, of about 10 to 15 degrees Centigrade, the difference between the readings of the thermometting and the soil surface thermometer was oscillating within the range of 5 to 7 degrees Centigrade.

If we accept the readings of the thermometting as the temperature of the soil surface, the above differences can be considered as corrections to the mercury temperature readings. The fact that the readings of the soil mercury thermometer substantially underestimate the soil surface temperature, is confirmed by the data collected by the observatory at Tbilisi (Figure 74).

Hourly observations over a long period of time for soil temperature at the surface and at depths were conducted by the observatory at Tbilisi. Figure 74 is a graphic presentation of the temperature variation with depth of the upper 10-centimeter layer of the soil, corresponding to the moment of absolute maximum at the soil surface, for each month from January to June 1890 -- in semilogarithmic coordinates.

Figure 74. The variation in the absolute monthly temperature maximum of the soil with depth (in semilogarithmic coordinates). Tbilisi, 1890.

In the layer from 10 to 1 centimeter, the vertical temperature variation, in the accepted coordinates is close to being rectilinear. This indicates that the maximum temperature variation with depth is in proportion to the logarithm of the depths. But in the layer from 1.0 to 0.1 centimeter (the location of the surface thermometer is conditionally adapted to the depth of 0.1 centimeter), isothermy practically prevails, and at times, as during the month of March, even an inverted temperature variation occurs. If we assume that the logarithmic regularity of the vertical soil temperature variation is maintained up to the top millimeter, then, in extrapolating along a straight line (see dotted line in Figure 74), we obtain for the top millimeter layer of the soil a temperature, which is by 10 to 15 degrees higher than the one recorded by the surface thermometer. This is in agreement with the data by L. I. Zubenok, since, during the soil surface absolute maximum temperatures, the difference between the latter and the air temperature at the height of 1.5 to 2 meters is greater than 20 or even 25 degrees Centigrade. Substituting this value into the right-hand part of formulas (39 and 40), we derive the correction to the reading of the mercury thermometer, which correction is approximately equal to 10 to 14 degrees Centigrade.

In addition, we can establish by Figure 74 that the readings of the surface thermometer correspond to the soil temperature at the depth of 0.5 to 1 centimeter. Observations at the Pavlovsk and Sverdlovsk Observatories confirm this.

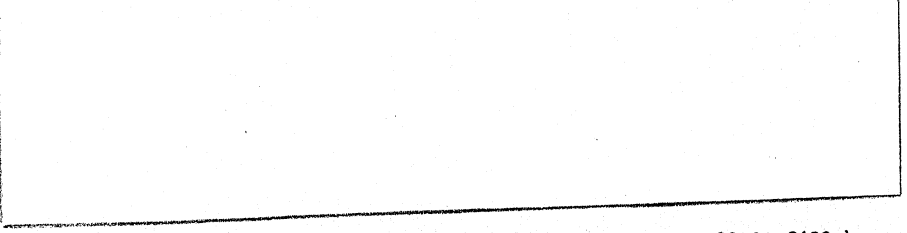
Thus, we arrive at the conclusion that soil surface thermometers record the temperature not of the surface itself but of the top layer of the soil to the depth of approximately one centimeter. The temperature of the surface itself will be higher than that during the day and lower than that during the night. Although the surface thermometers do not record the exact temperature of the surface itself, they are still characteristic of the thermal cycle of the very top layers of the soil, which is substantially different from the air temperatures, and which can, therefore, be utilized in microclimatic observations.

In addition to the above described thermometry used for the determination of the soil surface temperature, thermocouple junctions (devices by Pankevich, Chudnovskiy, and others) are used. However, the most perfect method for the determination of the temperature of the active surface is the radiational method, which is now coming into use.

Chapter 23

ORGANIZING AND CONDUCTING MICROCLIMATIC OBSERVATIONS
INCLUDING SAME IN THE MIDST OF HERBACEOUS VEGETATION

The microclimate of the upper layers of the soil and the adjacent air layers is linked directly to the strictly local characteristics of the soil and its vegetation cover, as a result of which it is distinguished by exceptional variability in space (see Figure 75).



0200 0400 0600 0800 1000 1200 1400 1600 1800 2000 2200 2400 hours

Figure 75. Soil temperature at the meteorological observation platform -- Sablino, LGU, 7 to 14 July 1938. A - 5-centimeter depth; B - 10-centimeter depth. 1 - denuded soil; 2 - natural cover; 3 - meadow trampled upon; 4 and 5 - meadow.

Student climatologists of the Leningrad State University conducted observations of soil temperature at six points of the exposed level platform of the meteorological station at Sablino (Leningrad), covered with meadow vegetation, which in spots was

trampled upon. One section was denuded of all vegetation. Included in the six observation points, was the basic thermometric installation of the meteorological station under a natural cover, which installation, it would appear, was to record the temperature of the soil under meadow vegetation. However, the natural cover of the station was a grass cover that became sparse on account of constantly conducted observations, and its height near the thermometric installation was only 5 to 8 centimeters. The soil under it was compact. As a result (see Figure 75), the temperature of the soil at the depth of 5 centimeters under the natural cover was practically the same as if the soil were denuded. During the day, the temperature was by 5 to 8 degrees higher than under the undisturbed grass covering the greater part of the meteorological platform to a height of 70 centimeters.

At the depth of 10 centimeters, the soil under the natural cover occupied an intermediate position between denuded soil and a soil with a dense grass cover, differing, however, from the latter, during all the observation days, by 3 to 5 degrees. Such variegation may be encountered at any meadow and in the field, not to mention the fact that we can cause it to appear ourselves, by disturbing the vegetation in the process of observations.

TABLE 114

The effect of disturbing a wheat grass cover upon
the air temperature. Agro-hydro-meteorological
Institute, 1939.

Date	Time of observation	Disturbed grass cover			Undisturbed grass cover		
		0 cm	2/3 of growth	150 cm	0 cm	2/3 of growth	150 cm
8/VIII	1428 - 1455 hours	24.4	23.6	23.6	21.9	23.2	23.8
9/VIII	1433 - 1510 hours	25.0	23.5	23.4	22.3	23.3	23.5

Table 114 above cites comparative data as between a platform with an undisturbed grass cover and a platform, on which the wheat grass cover was somewhat disturbed as a result of preceding observations. At the soil surface, the undisturbed grass cover caused during the two days a temperature lower by 2.2 and 2.7 degrees than at the platform with the grass cover trampled upon. The platforms even differed by the sign of their vertical gradient, i.e., in the midst of the disturbed grass cover there was a superadiabatic gradient, in the midst of the undisturbed grass cover -- a temperature inversion.

At the "bald spots" in the midst of a dense grass cover, higher temperatures occur than in the case of black fallow, since, in the presence of homogeneous conditions relative to the radiational balance, the interchange of air at the "bald spots" is diminished considerably as compared with the open sections of black fallow.

In studying the microclimate of cultivated vegetation, it is necessary to comply fully with the technique prevailing in agricultural production. The representativeness of the observation platform and its grass cover relative to production conditions must be the most important consideration not only in the selection of a platform, but also during the entire period of observations.

The microclimatic characteristics of any grass cover are evaluated by comparison with the microclimate of a denuded

soil surface (black fallow). Black fallow is the standard, the comparison with which permits the checking of observations, taken at different times, of the microclimate of various vegetational combinations and agricultural crops.

A black fallow section should always be near a section with a grass cover that is being studied. It is imperative that the two sections be uniform in all other conditions, except the grass cover (in the form of terrestrial relief, in the soil characteristics, in the degree of dampness), since only then the observed differences will be attributable to the presence of particular features in the grass cover. With the proximity of the two sections, the work entailed can be performed by one observer. At the same time, the minimum dimensions of the observation section, that will make the results representative and applicable to actual production conditions, are to be decided upon.

It is clear that no standard minimum observation area can be established. The size of the observation area will depend on the character of the surrounding terrain. The more the characteristics of the active surface of the surrounding terrain differ from the characteristics of the active surface of the area under study, the larger the latter is to be. For instance, in the comparative study of irrigated and non-irrigated sections of a wheat field, the size of each section is, of course, to be considerably larger than in the case of studying the climate of

a wheat field, when the latter is located amid sections of rye, barley and other crops, close to each other by their microclimatic features.

For the analysis of the effect upon the microclimate of the distance from the edge of the field toward its center, see Figure 76, constructed as per observations by E. S. Pankevich made at a kolkhoz field. With distance from the edge of the field, the temperature is diminished both inside the grass cover and above it.

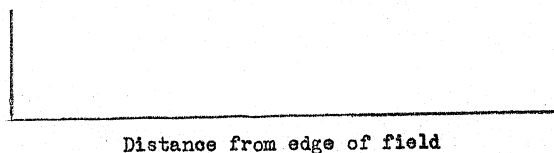


Figure 76. Variation in the air temperature with distance from the edge of a wheat field.

Ovtsino, AGMI, 23 July 1937.

Small-size black fallow sections distort the thermal cycle of the soil, due to the weakened turbulent heat exchange soil-air. A confirmation of this can be found in the data presented in Table 115, taken from the thesis by E. S. Pankevich.

TABLE 115

Variation in the temperature of demuded soil at the
depth of 5 centimeters in various size plots --
Ovtsino, 23 July 1936

Plot dimensions (in meters)	Hours														
	0700	0800	0900	1000	1100	1200	1300	1400	1500	1600	1700	1800	1900	2000	2100
1.5 x 1.5	19.2	20.3	22.6	24.4	26.0	27.5	28.4	29.1	29.0	28.5	27.9	26.6	25.3	24.1	22.8
2 x 2	19.0	20.2	22.5	24.0	25.5	27.2	28.0	28.5	28.7	28.2	27.9	26.8	25.6	24.5	23.2
8 x 10	19.8	20.2	21.9	23.1	24.5	25.8	26.8	27.3	27.6	27.3	27.1	26.6	25.6	25.0	23.3

In the midst of a meadow, with a grass cover about one meter high, three sets of platforms, with dimensions indicated in the Table, were denuded of all vegetation (altogether nine platforms). As seen from the Table, the mean value of the triple temperature reading for the soil at the 5-centimeter depth during the day was the higher, the smaller the platforms.

Considerable non-uniformity in the upper layers of the soil, mole holes, stone inclusions, rootstocks, microrelief, and non-uniformity in the grass cover result in cases when two practically close by thermometers at the 5-centimeter depth differ in their readings by 5 degrees or more. The previously described observations by N. N. Trankevich with 16 side-by-side thermometers at the 5-centimeter depth have shown that, on clear days, only quintuple readings provide for a precision of $\pm 1^{\circ}$. However, according to observations near Leningrad, already a triple reading at the 5-centimeter depth results in the same degree of precision. This variegation is, naturally, leveled off with further depth into the soil.

To avoid the distorting effect of microoscillations, calculations are to be based at least on triple readings. The usual procedure when observing air temperatures at various heights, is to use one psychrometer, which is shifted twice, or even thrice, in sequence (from lower layer to upper layer, from upper to lower, from lower to upper).

From all the above said, the following conclusion, important both from the practical as well as the theoretical angle, can be drawn.

The degree of precision in microclimatic observations, regardless of the preconceived notion of some climatologists, is in a number of cases below that in macroclimatic observations. In order to raise the degree of precision, it is not as much the use of finer instruments that is called for, as the practice of repeated set-ups and repeated instrument readings for each individual observation. However, even when this is done, we can only approximate the degree of precision attained in macroclimatic observations. It follows that the demand for higher precision in the special microclimatic apparatus is rather superfluous, with the exception of anemometric apparatus. With the non-uniformity of microclimate even within the limit of a single field, as analyzed in the preceding text, it is obvious that microclimatic analysis to the precision of even 0.1 of a degree has no realistic meaning.

In any microclimatic observation set-up, the proper selection of the air layer height and the soil layer depth is of great importance.

In selecting a height (depth) for the instrument set-up, it is necessary, on the one hand, to proceed from the characteristics of the vertical variation of the contemplated element, and, on the other hand, from the specific purpose of the assignment. For instance, in observations conducted for the purpose

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In selecting a height (depth) for the instrument set-up, it is necessary, on the one hand, to proceed from the characteristics of the vertical variation of the contemplated element, and, on the other hand, from the specific purpose of the assignment. For instance, in observations conducted for the purpose

of studying the ecology of vermin and pests, the instrument set-up is to be at the height (depth) of their maximum accumulation.

In the absence of a grass cover, temperature, humidity, and wind vary with height (in first approximation) in direct ratio to the logarithm of heights. The vertical gradients of the above elements diminish in direct ratio to height. It follows that the uniform distribution of observation points in height (as, for instance, at 50, 100, 150, and 200 centimeters) is improper. The temperature variation in the height interval between 50 and 100 centimeters is very insignificant, making it advisable to eliminate the 100-centimeter level setting. However, in the air layer up to 50 centimeters, it is necessary to add an intermediate instrument setting. The most important observation heights over denuded soil or over soil with a sparse grass cover are as follows: 5, 20, 50, and 150 centimeters. These heights have lately become standard, and all other observations are linked up with them. The standardization of these heights is determined by the fact that, at the 20-centimeter level, the radiational and ventilation errors of the ventilation psychrometer, in their mean values, cancel each other out. The 150-centimeter height is selected for the reason of being accessible for direct readings without the use of extra substructures and ladders, which already become necessary for observations at the height of 200-centimeters -- the standard height for general meteorological observations in the USSR. In addition, the tempera-

ture at the 150-centimeter level is practically no different from that at the 200-centimeter level. This is a direct result of the logarithmic regularity in the variation of meteorological elements with height in the lowest atmospheric layer, which is substantiated by voluminous empirical data (from the archives of the Department for Agricultural Meteorology of the All-Union Institute for Plant Cultivation), a part of which was cited in Table 116.

TABLE 116

Mean air temperature differentials at the height
of 150 - 200 centimeters as per observations
in Selyaninov booth. Tbilisi, 1932.

Months	Clear (0-2)			Cloudy (3-6)			Overcast (7-10)		
	Hours								
	7	13	21	7	13	21	7	13	21
V	0.1	0.0	-0.1	0.0	0.0	-0.1	—	—	0.0
VI	0.1	0.1	-0.1	-0.1	0.1	-0.1	-0.1	—	-0.2
VII	0.0	0.1	-0.1	0.1	0.1	-0.1	0.0	-0.1	0.0
VIII	0.1	0.2	-0.2	0.0	0.1	-0.2	0.0	—	-0.2
IX	0.0	0.2	-0.2	-0.1	0.1	-0.2	0.0	0.1	-0.1
X	-0.1	0.1	-0.2	0.1	0.1	-0.1	0.0	0.0	-0.1
XI	0.0	0.1	0.0	0.0	0.1	0.0	0.0	0.0	0.0

The selection of observation heights in the midst of a grass cover is more complex. In this case, the observations are adapted to those heights, where a pronounced change in the variation of the contemplated meteorological element may be expected, with the upper observation point remaining at the height of 150 centimeters. The break in the vertical air temperature distribution curve is adapted to the surface of the grass cover and to a height corresponding to two thirds of its growth. Lower down, in addition to the 20-centimeter level, the observation height of 5 centimeters is added.

Since the vertical temperature variations in the soil, which are very abrupt in the top layer, become diffused with depth, the vertical intervals between observation points in the soil must become greater with greater depth. Usually, thermometers are set at 0.5- and 15-centimeter depths. However, for the computation of the thermal balance of the soil, it becomes necessary to obtain additional data pertaining to the 10- and 20-centimeter depths.

Observation periods. Systematic observations are limited to 2 - 4 periods. Episodically, round-the-clock observations at one- or two-hour intervals are conducted.

In selecting observation periods, it is necessary to remember that, during the diurnal period, the algebraic sign of the vertical air temperature gradient changes. The time of the transition from the daylight superadiabatic gradients

to the nocturnal temperature inversion varies basically in relation to latitude and time of year (see Table 8).

The nature of the vertical temperature gradient determines the entire meteorological complex. With the setting in of the temperature inversion, turbulent exchange is sharply diminished, evaporation practically ceases, up to the point when it is reversed into condensation (dew, hoarfrost), wind velocity is reduced.

During the day, the maximum gradient occurs about midday (1200 to 1300 hours). Observations are to be adapted specifically to this moment, particularly since it coincides with the moment of general meteorological significance.

The selection of the time for nocturnal observations is somewhat difficult, since there is no moment of maximum temperature inversion value. In those cases when the maximum interest is concentrated on the vertical gradient, the observations are conducted at 0100 hours of the night. The advantages offered by this hour, as well as by the 1300 hours, are that they coincide with the hours of general meteorological significance. In the case when maximum interest is concentrated on the temperature minimum and its accompanying conditions, observations are to be conducted directly before sunrise. The technical advantage, available at this hour, is that ~~no~~ artificial lighting is ~~no~~ longer necessary. The general meteorological hours, 0700 and 1900, are not suitable for observations in the surface air

layer, since, during the greater part of the year they are timed to the transition moments of disintegration and evolution, respectively, of the temperature inversion, and are distinguished by greater changeability in the variation of almost all meteorological elements.

For the analysis of the diurnal variation of meteorological elements, it is necessary to conduct episodic round-the-clock observations at 2-hour intervals.

In the case when the observer is in charge of two points (the basic point and the check point), readings are taken first at the basic point, then at the check point, and again at the basic point.

If the basic problem is the study of the thermal cycle of the soil under field conditions, when observation points are distant from each other, and the observer cannot take all the readings at the specific moments, at which they are recorded, the necessity arises for shifting the moments of observation to the periods of minimum temperature change, i.e., to the periods of temperature maxima and minima.

The shift of observation time to the above periods makes it possible, on the one hand, without any particular damage to precision, to stretch the observations over a relatively greater period of time (up to one hour or even longer), and, on the other hand, it provides for the comparative study of

the thermal cycle at the most actual moments of temperature extremes.

Table 117, compiled as a result of round-the-clock observations at two-hour intervals during the sunny summer days near Leningrad (by the Leningrad State University Meteorological Station), under the column title "Hours", shows periods, during which the temperature differs from its maximum and minimum values by not more than 0.3 of a degree Centigrade. The greater the depth, the greater, naturally, is this period of time. In general, by observations at 0600 - 0700 hours all the variants at all depths were included in a temperature value close to the minimum, and only the temperature of the soil at the 20-centimeter depth, under a meadow that was mowed down, was at this time by 0.4 degree above the minimum value.

Observations at 1600 - 1700 hours show soil temperatures close to its maximum value, with the exception of the 20-centimeter depth, since at this depth, at the above indicated hour, the temperature differs from its maximum value by 0.3 - 0.6 percent, but, then, it is very close to the mean diurnal (difference 0.0 - 0.3 degrees).

With the above indicated observation periods, the mean diurnal temperature is computed by the formula $\frac{\text{max} + \text{min}}{2}$. The divergence with the mean diurnal temperature value, computed by round-the-clock observations, as seen from Table 117, does not exceed, in its mean value, 0.1 - 0.2 degrees (only in one case it was 0.3 degree).

TABLE 117

Mean soil temperatures under various types of grass cover
over period from 7 to 14 July 1938, Sablino

Character of vegetation cover	Depth in centimeters	Mean diurnal value		Mean maxi- mum value	Hours	Mean mini- mum value	Hours
		From 12 round- the-clock ob- servations	$\frac{\text{max} + \text{min}}{2}$				
Demuded soil	5	19.7	19.8	23.8	1600	15.7	0400-0600
	10	19.5	19.6	22.9	1600-1800	16.2	0400-0600
Natural cover	5	19.5	19.7	23.2	1400-1600	16.1	0400-0600
	10	18.3	18.4	20.0	1600-2000	16.6	0600-0800
	20	17.4	17.5	18.3	1800-2400	16.6	0600-1200
Clover	5	15.6	15.5	17.0	1600-1800	13.9	0400-0600
	10	15.3	15.3	16.2	1600-2200	14.3	0400-0800
	20	14.7	14.7	15.1	1600-2400	14.2	0600-1400
Mowed meadow	5	19.8	20.1	23.2	1600	16.9	0400-0600
	10	19.3	19.4	21.4	1600-1800	17.4	0600-0800
	20	18.2	18.1	18.8	1800-2400	17.4	0800-1200
Grass-leguminous meadow	5	16.0	16.0	17.2	1400-1800	14.4	0200-0600
	10	15.6	15.6	16.4	1600-2200	14.8	0400-0800
	20	15.2	15.2	15.5	1600-2[?]	14.8	0600-1200

A similar analysis of the diurnal temperature variation of the soil was made by N. N. Trankevich by the data of the Far East Agro-Meteorological Station. His conclusions relative to two-time observations (at 0800 and 1600 hours) coincide with the above tabulated. Experience in dealing with peat soils indicates that, due to the lag in them of the diurnal temperature variation, both observation times have to be shifted by one hour.

Since for the obtaining of representative data relative to air temperature, single readings are inadequate, two or even three readings at each height are usually taken. Ventilation psychrometers can be set up simultaneously at all observation heights. However, with the scarcity in instruments, with the necessity for parallel observations at the check areas, and the advisability of obtaining double readings kept in mind, the same psychrometer is usually utilized at two or even four heights, by shifting it in sequence from the upper layer downward two or three times.

Such a method of observations even has some advantages by introducing the equalization of the systematic errors.

Prior to observation time, the psychrometer (the entire device) is to have a temperature close to the air temperature, which is usually attained by exposing the psychrometer in the open air for 20 minutes. It is to be remembered, however, that being left in the sun, without ventilation, the psychrometer

can become considerably overheated. In such a case, it is to be ventilated for 10 minutes prior to use. A reliable index for the ventilation having been adequate, is the absence of stable changes in some definite direction. However, the oscillations of temperature about a certain level is a natural phenomenon, which can be eliminated only at the expense of increasing the reluctance of the device. The water for sprinkling is also to be of approximately the same temperature as the surrounding air. When all the above instructions are observed, the psychrometer can be used for observations three minutes after being sprinkled and set. The other readings can be taken at intervals of 1 to 1.5 minutes. One sprinkling is effective for 8 to 10 minutes during the day, and for 20 minutes during the night. After each reading, the psychrometer is to be reset, since after the third minute the rotation velocity of the ventilator begins to abate.

When the check platform is sufficiently close to the basic observation platform, the observer, while taking down the reading at one platform, sets the device at the next height, and proceeds ~~himself~~ to the check platform to take the check reading.

The temperature of the soil is tabulated either during the intervals between the psychrometer observations or after they have been completed.

For the physical substantiation of some of the characteristics of the microclimate, even though from the qualitative angle only, all microclimatic observations are accompanied by a description of the degree of sky cover, wind, and moisture content of the active surface at the time of the observations.

The sky cover analysis together with the date and the hour of the observations characterize the approximate radiational balance of the active surface. By the wind analysis, we can judge indirectly the value of the turbulent exchange, and the evaluation of the moisture content of the active surface, inclusive of the hydrometeors, furnishes the concept of the possible losses of heat to evaporation.

The direction of the wind, cloudiness (general formations and lower formations), and the hydrometeors are determined ahead of and after the instrument reading tabulations. To determine wind velocity, a hand-operated anemometer is switched in for a period of 10 minutes, conforming to the time of taking the ventilation psychrometer readings. A notation is made of the actual moment the sun rises for the observation point involved. In addition, during the day, with sky cover variability, special designations are used as follows: (☉) - for clear weather; (☁) - for overcast; (☉|☁) - for the sun visible from behind the clouds. The latter notations permit the evaluation of each observation from the angle of being typical for the specific conditions of weather.

Chapter 24

METHODS FOR THE STUDY OF LOCAL CLIMATE - ANEMOMETRIC
AND THERMOMETRIC SURVEY

Stationary and field observation methods are employed in the study of local climate. Field observations and stationary observations, which are tied in directly with field observations, are combined under the general concept "survey" (thermometric, anemometric, snow-gauging, etc).

There are special instructions relating to snow-gauging surveys, and we will, therefore, limit ourselves here to the analysis of the anemometric and thermometric survey.

Anemometric Survey

The purpose of anemometric survey is to analyze the deformation of the basic air current and the formation of local wind in relation to the presence of interlaced relief, bodies of water, vegetation, including special forest plantings (field-protecting forestation strips), and various structures.

The initial undisturbed current is observed at some height over the terrestrial surface (200 to 500 meters). At the surface, it undergoes deformation in relation to the characteristics of the underlying surface at the given place. A certain part is played by the preceding deformation of the current, and, consequently, by the character of the underlying surface along its path.

If in classifying the types of terrestrial relief, the character of a locality lying along the path of the motion of the air at a given point is considered, i.e., if for example, windward and leeward slopes are differentiated, it means that the history of the air current is partly taken into account. The initial current itself is not accessible to the conventional methods of surface observations, for which reason the wind at the same heights as the points of observation, but over a specially selected exposed level place, is used as its conditional exponent.

For the quantitative analysis of the deformation of the current, the ratios of wind velocity at the observation point to the wind velocity at an exposed level place are used, and for the direction -- the corresponding differences of the bearings (in a 16-rhumb scale).

Since both the formation of local winds and the deformation of the basic air current are in relation to the degree of atmospheric stability and of the turbulent exchange, determined by the latter, both phenomena, must depend on the weather, time of the year, and hour of the diurnal period.

The deformation of the air current is approximately in an inverse ratio to the logarithm of the heights. At its maximum value below, this deformation is extinguished with height.

All the above enumerated characteristics of the wind cycle are to be taken into account when organizing an anemometric survey.

For wind velocities beginning with 1 meter per second and above, a hand-operated anemometer is completely suitable. When necessary, it can be replaced with a Tret'yakov wind gage.

Most of the wind velocity observations in the lowest atmospheric layer have been and are being made with the aid of hand-operated anemometers. The principle of its design and its appearance are described in the thesis by V. K. Kedrolivanskiy. The anemometer meter may be engaged and disengaged from a distance of several meters, which permits its mounting on poles up to 5 meters and even more in height. The anemometer is calibrated, beginning with one graduation per second. Converted into terms of meters per second, this will show wind velocities beginning with 1.0 to 1.4 meters per second. This specifically is one of the basic deficiencies of the device, which practically prohibits its use for heights below one meter. Even at the height of one meter, at some points, the nocturnal wind velocities could not be gauged by a hand-operated anemometer in 50 percent of all cases.

The conversion factor of the anemometer has a precision of 0.01, and, consequently, officially provides for wind velocity computations ranging up to 10 meters per second with a degree of precision equal to 0.1 meters per second.

Practically, the anemometric errors under field conditions are considerably higher. Part of the errors is linked to the features of the design itself. As all anemometers with the Robinson cross, the hand-operated anemometer records some mean

leveled-off velocity value, which is usually greater than the mean arithmetical value, the difference between the two being the greater, the more frequent its pulsation and the greater its range. With a velocity of 1 meter per second, when the wind, in abating, attains a value below the sensitivity point of the anemometer, and the anemometer comes to a complete stop, the pulsation, on the contrary, can induce a recording, which is below the mean velocity value. In the presence of mild winds, the above may result in the systematic exaggerations of the vertical wind gradients, due to the above circumstance predominantly affecting the lower level anemometer, where the wind velocity is lower. Due to the proximity of the meter to the perceptive part of the device, its recordings are in relation to the location of the frame with reference to the direction of the wind. It follows that, in order to eliminate superfluous errors, the anemometer is to be mounted in such a way that its meter is normal to the wind, and, specifically in this position, the anemometer is calibrated.

All errors inherent in the characteristics of instrument design may be considered, in the first approximation, to be constant. While affecting to the fullest degree the absolute values of the mensuration results, these errors may be, to a certain extent, leveled off by the method of comparative evaluation of ratios and differentials, but only on condition that all investigations are performed by devices of the same type. This is constantly to be taken into account in observational procedure.

A considerable part of the observational errors, resulting from the use of hand-operated anemometers, is due to the non-stability of the calibration curve. This calls for systematic checking of the anemometer under actual field conditions.

In observing wind velocities by hand-operated anemometers, the direction of the wind is determined by the pennant, which is a narrow strip of white cambric, 70 centimeters long, fastened to a 2-meter pole. For the convenience of reading the wind direction, eight bearing-indicating pegs are set around the pennant pole before observations begin. The bearings are distributed with magnetic declination taken into account. If the magnetic declination is in excess of 5 degrees, a corresponding mark is made on the compass (mark is made on the graduation, with which the magnetic needle is to coincide when the orientation of the compass points is correct).

The Tret'yakov wind gage, widely used at the present time, shows the direction and velocity of the wind, with the latter indicated by the deflection of a metal blade from a vertical position.

The degree of precision of the wind gage is 0.5 meters per second within the velocity range of 1 to 6 meters per second, and 1 meter per second at higher velocities. This wind gage is frequently used alongside the hand-operated anemometer for the determination of wind direction.

In planning observations, it is necessary to differentiate between general and specific anemometric surveys. The first requires for its execution a period of no less than 6 months, and it must embrace, at least partially, the warm and the cold seasons of the year. It must be conducted by day and by night, under various weather conditions and at various heights over the surface of the soil, since all the above stipulations exert a direct effect upon the deformation of the air current (see Chapter 12).

A specific anemometric survey has the assignment of determining the deformation of the air current at a predetermined height and under predetermined weather conditions. For instance, the determination of the foehn wind deformation, with reference to inquiries from the horticulture industry, calls for a specific anemometric survey only. In that case, observations are adapted to a predetermined height (the height of the fruit-bearing trees), predetermined weather conditions, and predetermined season of the year.

In addition, it is necessary to differentiate between the anemometric survey of specific parcels, to which parcels alone the results of the survey will be applicable, and the survey to determine the deformation regularities of an air current in relation to the characteristics of a locality, for the purpose of applying the results to any areas with similar forms of terrestrial configuration, vegetation, or building construction.

If, in the first case, the clarification of the part played by individual components of local conditions (relief, vegetation), may be of practical interest, bearing in mind the possibility of impending change in some of them (for instance, the clearing of an edge of a reforestation planting), certainly in the second case, the accountability of the part, played by each of the components of the complex, acquires a particular importance.

Having determined the effect of the individual elements of the complex, as, for instance, the width or the height of the forestation strips, it becomes possible, subsequently, with the aid of the effect characteristics of the individual elements to pass on to the evaluation of the combined effect of more than one element, as for instance the effect exerted by forestation strips of a given width and height.

However, complexes of special interest, as, for instance, the width and the wind penetrability of the field-protecting forestation strips, are to be taken into account as a whole, since their total effect is, by far, not in all cases, merely the algebraic sum of the effects of each individual component.

A series of practical instructions in the conducting of an anemometric survey is given in the Appendix.

Thermometric Survey for the Determination of the Frost Hazard of
an Area

Of all the local characteristics of the temperature cycle of a territory, the maximum practical importance is attached to the frost-hazard characteristic. The evaluation of the frost hazard is of the utmost importance not only to agriculture, but also to other branches of the national economy, since these frost areas are at the same time distinguished by maximum diurnal temperature ranges, by low relative humidity of the air, and by nocturnal air stagnancy.

The basic assignment of the thermometric survey is to determine the degree by which the frost hazard of the contemplated parcel is higher or lower than that of the nearest meteorological station. A final quantitative evaluation can be had only when a corresponding juxtaposition with observations at a point, where the contemplated climatic characteristic has been established on the basis of perennial observations, is available.

A thermometric survey is effected with the aid stationary and field observations, both methods being intermittent in nature and conducted only during the time of the year, when the frost hazard of an area is of practical interest. In the subtropical zone, observations are conducted in the winter (from November to March inclusively), in the north -- during the summer, in the central belt of the USSR -- during the spring - autumn months.

A network of stationary thermometric points is situated either at the basic types of locations, which are characteristic to a given climatically homogenous zone, or at parcels having a direct and immediate interest (such as a cotton field, a tea plantation), regardless of their typicalness for the surrounding territory. In both cases, the organizational center for the thermometric network is the nearest meteorological station.

When selecting an observation locality, two-three preliminary field thermometric surveys are to be made (the method to be used will be described later in the text), in order to eliminate the chance of the selected points representing typical conditions.

The basic observations at the thermometric points are made at a height of 1.5 to 2 meters in conventional or simplified type observation booths. Additional observations may be made at the very lowest air layer and on the soil surface. The adaptation of the basic observations to the height of 2 or 1.5 meters, inasmuch as the temperature at both these heights is practically the same, is determined by the necessity of linking them up with the basic meteorological network. In addition, the temperature in the very lowest air layer depends on the strictly local characteristics of the active surface, it is an element of microclimate that can be changed with relative ease in the development of the parcel. In studying the frost hazard, we are interested in the more stable features of the local climate, which are characteristic of parcels several tens and even hundreds of meters in extent. The

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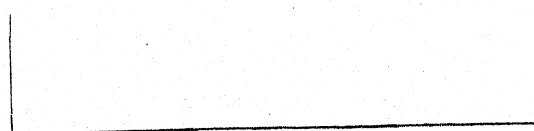
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frost hazard characteristics of such areas are manifested at the height of 2 meters. The question, naturally, arises, as to what degree the difference in the frost hazard at the height of 2 meters will serve as an index for the frost hazard at the height of the grass cover. A reminder is in place that the difference in frost hazard value is determined basically by the flow off and the stagnancy in the radiationally chilled air.

Let us analyze 3 extreme cases. At an exposed level place there is no flow off and no inflow of cold air, and there occurs, so to speak, a normal temperature inversion. On a hill, where the chilled air will flow off, with the warm air from the free atmosphere flowing in its place, the deflection of the thermal stratification from the normal will be determined by two factors: the advection itself of the warm air will increase the value of the temperature inversion, and the accelerated exchange, on the contrary, will decrease its value. In the presence of negative terrestrial configuration, the process is inverted: the advection of the cold air decelerates the temperature inversion, and a weakened exchange -- accelerates it.

In the final analysis, due to the mutually compensatory effects of advection and turbulent exchange, it would be reasonable to expect that the vertical temperature stratifications of parcels with various degrees of frost hazard will differ little from each other. In such a case, the difference in the minimum temperature values for parcels with a different degree of frost hazard will be of the same order of magnitude at all heights.

This is confirmed by direct observations. Figure 77 is a graphic presentation of the temperature differences (the departure from the temperature at the center of the slope), as per observations at the heights of 1.5 and 0.5 of a meter over the hill slopes of the Ureki Sovkhoz (Western Gruzija). The difference in heights between the extreme points is 40 to 50 meters, the distance between the points is 10 meters. The departure from the temperature at the center of the slope for the heights of 1.5 and 0.5 meters is of the same order of magnitude, but the temperature variation is more gradual at the height of 1.5 meters. All this confirms the representativeness of the frost hazard evaluation at the height of 150 centimeters for the entire surface air layer. Additional observations in the very lowest part of the surface air layer are called for only in those cases, when the parcel is already overgrown with vegetation that has economic value.



Points

Figure 77. Temperature of the air before sunrise at various heights over the surface of the slope -- in departures from the temperature at the center of the slope -- Sovkhoz Ureki, 1934. A -- observation route 14/II; B -- same for 27/II; B -- same for 7/II. 1 -- 150 centimeters; 2 -- 50 centimeters.

Stationary observations are made with the aid of minimum thermometers, with check observations by standard thermometers, both mounted in a simplified observation booth designed by Selyaninov (Figure 78). This booth is smaller, less expensive and easier to fabricate than the psychrometric booth. Due to its small size, it can be set up atop a single pole. It is usually placed at a height of 1.5 meters so as to eliminate the necessity for a substructural ladder for the observer. Special investigations indicated that the diurnal minimums, by observations in the Selyaninov booth at a height of 1.5 meters, corresponds to the minimums observed in the conventional booth within 0.2 degree. In the absence of booths, the minimum thermometer may be set up at the same height under a simplified Borisov-designed protection (see Figure 79). In that case, the diurnal minimum values will be somewhat underestimated as compared with the values obtained in a conventional booth. It will then become necessary to effect a similar setting at the meteorological station, which is involved in this thermometric survey, in order to render the results of the observations comparable.

Figure 78. Simplified meteorological booth for housing minimum and maximum thermometers, as designed by G. T. Selyaninov.

A -- side view; Б -- front elevation; В -- thermometer base; В₁ -- swivel-operated board covering the holes provided for three thermometers; Г -- detail of prop with grooves for shutters.

of the night to follow. When two-time observations are made, the observer makes the rounds of the observation points twice: in the morning and in the evening. The first observation is made at a time, which is possibly close to the actual temperature minimum, in order that the reading of the standard thermometer serves, to a certain extent, as a check, upon the recorded minimum temperature value. In addition, the first observation is to judge the degree of sky cover and wind at this critical moment, and also to record the presence of hoarfrost and damage to plants caused by freezeover. The purpose of the second observation is to reset the minimum thermometer for the following night (to move the pin). One of the advantages offered by two-time observations is the greater precision in the determination of the additional correction to the minimum thermometer reading, which correction is computed by values close to the minimum value.

Field temperature observations at certain point of a pre-selected route are an addition to the network of stationary point observations. Their basic purpose is to determine the temperature variations between the stationary points. In addition, they play the part of a preliminary reconnaissance, particularly, in the selection of spots for the stationary observations.

With relation to the method of transportation, thermometric surveys can be classified as pedestrian, automotive, and equestrian.

The possibility of field observations relating to the distribution of the minimum temperatures is based on the fact that, in the pre-morning hours, 1.5 to 2 hours before sunrise, the temperature varies very little in time. As is known, the temperature oscillations at this time are also very small. To check the possibility of utilizing the results of field observations, F. F. Davitaya juxtaposed the results of one-time and field (vari-timed) observations in January and February 1932, relating to four points of one of the survey routes in the environs of the City of Makharadze (Western Gruzija). The first observation point was on top of a very steep slope, the second -- on a hillocks in a valley, the third -- at the bottom of the valley, the fourth -- on an exposed hill.

Table 118 cites the mean-square difference between the results of field observations (π) and one-time ($\circ$) observations over a 9-day survey period.

Table 118

Mean quadratic difference between field (vari-timed) observations and one-time observations by the formula

$$\Delta = \sqrt{\frac{(\pi - \circ)^2}{n - 1}}$$

	Points of observation			
	1	2	3	4
Δ	0.42	0.31	0.21	0.26

The mean quadratic difference in all cases is less than 0.5 degree. The minimum error, as was to be expected, pertains to the valley bottom, where, due to air stagnation, the temperature oscillates to an insignificant extent only.

Remembering that the differences in the temperature of individual points during the survey are usually in excess of 3 to 4 degrees, we conclude that the error, due to the non-synchronousness of the observations may be disregarded.

In order to further reduce the error, due to the non-synchronousness of the observations at each point, the temperature is read twice, when arriving at a point and when leaving the point. In connection with this, the length of the route is so adapted that it is possible to walk it (or to ride it) both ways within 1.5 hours, inclusive of observation time.

The route is to pass through two extreme (by the frost hazard criterion) stationary observation points, otherwise the climatic utilization of the results of the survey (see Chapter 26) will be extremely difficult. Individual points of the route (they may number 15) are so selected as to facilitate to the greatest possible extent the interpolation of the data, pertaining to the stationary points, onto the surrounding terrain.

Figure 80 shows a diagrammatic distribution of stationary and field observation points under conditions of interlaced terrestrial relief and close to a body of water.

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Figure 80. Diagram of a distribution of observation points and routes in studying the distribution of the minimum temperatures.

1 -- meteorological stations; 2 -- meteorological stationary observation points; 3 -- field observation routes.

As already indicated, field observations are made in the pre-morning hours, beginning one and a half hours before sunrise, i.e. before the temperature begins to rise. Within the normal diurnal variation, the temperature at this time is at its minimum value, and it is stable.

Clear nights, with a wind velocity in exposed places of no less than 3 meters per second, are selected for observations. The presence of a high degree of cloudiness (Ci, Cist, Ast) does not hinder the survey procedure. It is particularly important to make these surveys on days, complying with the above indicated weather specifications during the arrival of cold waves, which, do not even result in frost. It being the case that it is important to make the survey not only on the day with the lowest temperature value at the meteorological base station, but also on the days that follow, when frequently the temperature in the negative forms of relief (hollows, synclines, etc.) drops to its lowest value.

Observations are made with the aid of a ventilation psychrometer, which, in extreme cases, can be substituted by a sling-type

thermometer. The psychrometer is suspended at the height of 150 centimeters (to the bottom of the instrument), on a tripod, or on a special pole.

It is recommended, when time permits, that supplementary observations for points of the greatest importance are made at the height of 20 centimeters. At this height, the psychrometer is suspended in a horizontal position. Alongside of this, recordings are made of the degree of cloudiness, the wind (sensory observation permissible), and the hydrometeors. The time of each reading must be recorded. The time of sunrise and the time the solar rays become incident upon the observation parcel must be recorded with particular meticulousness. When a sling thermometer is used, it is swung overhead until a stable temperature reading is attained.

For reconnaissance observations embracing a large area, automotive thermometric surveys are indispensable. One of the results of such a survey is presented in Figure 57.

An automotive thermometric survey was conducted by the Agro-Hydro-Meteorological Institute, under the supervision of G. T. Selyaninov. For this occasion, M. G. Gol'tsman and A. A. Znamenskiy designed a special photo-recording thermometer, which permitted the taking of observations without stopping the car (diagram presented in Figure 81).

Figure 81. Diagram of the photo-recording thermometer designed by Gol'tsman and Znamenskiy.

The photo-recording thermometer is a conventional mercury thermometer mounted integral with a film photo-apparatus. The device consists of three basic parts: a box containing the film, a camera obscura with objective lens, and a thermometer with an illumination device. The film-containing box is braced hermetically with the camera (cylindrical in shape), at the other end of which, also hermetically braced, is the illumination device and the thermometer. A high-lumen objective lens with a focal distance of 50 millimeters is mounted in the camera. The bottom part of the thermometer, the bulb, is shielded from possible injury by a fine metal grid. The thermometer scale illuminator consists of three electric bulbs, screwed into a reflector. The electric bulbs are fed from a 4 to 5 volt dry cell battery.

The entire device is shock absorber-mounted on the car, in front of the radiator. Observations are conducted along a predetermined route on the run. In passing representative parcels of the route, the illumination is switched on, and the scale and the capillary of the thermometer photographed. Figure 82 shows a sample of the tape of the photo-recording thermometer. Each snapshot occupies 5 millimeters of the film, and, on a film of standard size, about 100 snapshot recordings of the thermometer can be made.

Chapter 25

THE ANALYSIS OF PHYSICO-GEOGRAPHICAL FACTORS --

PHYTOINDICATORS

In studying the local climate and microclimate, it is important to record an exhaustive description of all the determinant factors, such as: local characteristics of terrestrial relief, vegetation, soil covering, proximity to bodies of water, the extent of structural development, and the like.

The characterization of the above enumerated physico-geographical factors is necessary not only for the understanding of the causes and the physical essence of the phenomenon under study, but also for the extrapolation of the conclusions drawn from the study onto other territories. In connection with the latter, the climatologist must be capable not only of describing physico-geographical factors per se but also of classifying them into types, by the nature of their effect upon the local climate and microclimate. Examples of such classifications are given in Chapters 12 and 13.

The characterization of physico-geographical factors may be descriptive (qualitative and quantitative) and graphical (diagrams, profiles, plans, maps, photos).

Objectivity notwithstanding, the characterization of a locality is done with due consideration given to the regularities of its effect upon the local climate and microclimate. These

regularities of effect serve as the criterions for the selection from the complex of natural phenomena of what is most essential. This is why the characterization of a locality can be undertaken only with the direct participation of the climatologist in charge.

The characterization must lay no claim to being all-embracing. Thus, in the thermometric survey to establish the minimum temperatures, basic attention is to be concentrated upon those features of the terrestrial relief and vegetation masses, which determine the flow off or stagnancy of the cold air during the hours of the temperature inversion. In studying soil temperatures, attention is to be concentrated upon the exhaustive characterization of the vegetation cover and the moisture content of the soil.

A description diagram for observation point localities embraces the characteristics of terrestrial relief, vegetation, and soil. In each concrete case, depending on the research assignment, some divisions of the above-mentioned diagram, are to be concentrated upon to the fullest possible degree, other divisions, on the contrary, may be disregarded.

Particular attention is to be concentrated upon the dynamics of the vegetation and the soil cover, also upon the boundaries of bodies of water (flood areas and areas of vernal desiccation). The description of a phenomenon, therefore, must not be limited by adaptation to a particular moment of time. It must be supplemented by the characterization of the seasonal

variation of the phenomenon. A diagram of such a description is given in Appendix 2.

The height of a locality is most frequently determined by barometric leveling.

The topographic profile is determined by leveling with three (or two) leveling rods and a level, or a plumb bob. The rods and the plumb bob may be made in the field. Figure 83 shows a diagram of leveling with two or with three rods.

Figure 83. Diagram of leveling with two or with three rods.

To determine the steepness of the slope, it is most convenient to use the acclivometer "Gedorazvedka" (Geological prospecting), with the leveling data utilized for this purpose. The acclivometer is also useful in determining the height of trees and structures.

The height of a tree h may be computed by formula:

$$h = r \tan \alpha + a, \quad (35)$$

where r is the distance to the tree; α is the angle of sighting the treetop; a is the level of the observer's eye.

To determine distances, it is best to use the two-meter rod ("sazhenka").

As a graphic description of the features of a locality, plot plans in various scales can be utilized. In addition to horizontal areas, they are to indicate vegetation, buildings, and instrument setups.

Under interlaced conditions of a locality a visual concept of the terrestrial configuration can be had by drawing a topographic profile map, on which vegetation and buildings are indicated by conventional designations, with the observation of the vertical scale.

In the presence of vegetation and structures, the construction of a topographic profile map is recommended even in the case of a level location. Profiles may be plotted along a radius from the observation point, as a center, in the direction of the prevailing winds.

The extent of the limitation of the horizon merits particular attention. The extent of limitation (in degrees) is best determined with the aid of a theodolite. A compass may be utilized for the determination of the azimuths, and an inclinometer for the determination of the height of the horizon. An example of graphic presentation of the height of the horizon was given in Figure 4. It is desirable to include in the same diagram the path described by the sun on the days of the equinox and the days of the summer and winter solstice, or for the period of observations. These lines in the diagram will permit the calculation of the lag in the actual rising of the sun and

its actual early setting due to the degree of limitation of the horizon of a locality.

In characterizing vegetation, it is useful to take into account the following features of its effect upon the climate. Plants, in shading the surface of the soil, exert an effect upon the radiational heating during the day and the radiational chilling during the night. Plants constitute an obstacle to the movement of the air, reduce wind velocity, induce the formation of vorticity, retard the cold air flow off, weaken turbulent exchange. And finally, the transpirational capacity of the plants, which varies in relation to the species of the plant, or even to its ecotype, phase of growth, and the quantity of its green mass, is of great importance. These specifically are the angles from which the vegetation analysis is to be approached.

A few words about the green vegetation mass. The vegetation mass characterizes not only the intensity of transpiration, but, in conjunction with the height of the plants, it determines the degree of covering the soil and the insulation of its surface from radiational heating and chilling.

The weight of the green mass of a grass cover can be determined as follows. A section is divided into half-meter plots, from which the grass cover is completely cut away with shears and immediately weighed. The values obtained are averaged and multiplied by two in order to arrive at the weight of the green mass per 1 square meter.

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Table 119 is an example of such mensuration (under various conditions and the use of various agro-techniques) for a number of points in the USSR.

Table 119

Weight of the green mass of wheat
per 1 square meter (in grams)

Phases of growth	Number of cases	Maximum	Minimum	Under arid conditions	Under adequate moisture conditions	Sorts of wheat	
						Cesium	Lutescence
							062
Fascicular	57	326	24	--	--	102	104
Spiking	44	1822	116	312	812	660	596
Ceraceous maturity	50	1880	100	430	708	642	568

The Table shows the extent to which the weight of the green mass of wheat may vary, i. e., the extent to which it will shade the soil and the amount of moisture and heat consumed for transpiration.

In addition, vegetation may be utilized as a unique climatic phytoindicator.

The method of phytoclimatic indicators finds practical application in the agricultural penetration of new territories, when the direct meteorological data available are very limited in scope or completely absent. Under these conditions, data on the presence of vegetation, which shows a negative reaction to certain climatic characteristics, as, for instance, perishing from a frost of predetermined intensity, will prove the absence of such characteristics (in this case -- the absence of destructive frosts) in the new territory.

This method is also applicable to the study of local climate and microclimate. In 1934, the phytoindicator method was used by the Agro-Hydro-Meteorological Institute in the agro-climatic zoning of the subtropical belt, the boundaries of which are stipulated by winter frosts of predetermined intensity.

A special grouping of the phytoindicators of the subtropical climate was made by P. S. Romanovskiy, as cited in Table 120.

Table 120

A simplified grouping of phytoclimatic indicators characteristic of the three basic zones of the subtropical belt of the USSR

Groups	Name of zone	Phytoindicators	Mean values of the annual
			absolute minima (in degrees)
I	Citrus fruits	All species of citrus fruit trees, palms (pritchard, date, cocoanut), eucalyptuses, Australian acacias, agaves, club palms, tung trees, oleanders (trees and high underbrush)	From -2 to -6
II	Tea and olive	Laurel, sweet bay (Laurus nobilis), camphor laurel, hemp palm, camellia, strawberry tree, low underbrush oleander	From -7 to -8
III	Extreme boundary of subtropical zone	Fig tree, Japanese persimmon, pomegranate tree, cypress tree	From -9 to -11

The last column of Table 120 cites the mean values of the annual absolute temperature minima, which are characteristic for each of the three zones. The local climatic features of individual microzones at the lower slopes of the Talysh mountains, in the area of Lenkoran' and Astara, were evaluated with the aid of the above Table. Subsequent meteorological observations substantiated the correctness of this microzoning.

Phytoindicators, as an indirect method for the characterization of a climate, require a thorough analysis of the entire complex of conditions in order to avoid errors of judgment. Particular discretion should be exercised when using cultivated plants as phytoindicators, since they may be artificially shielded from the pernicious climatic conditions, and, so to speak, exist in spite of the latter. It is also necessary to take into account the particular set of conditions, to which the presence of a certain type of vegetation is attributable: whether to features of local climate, embracing relatively large areas, or to microclimatic characteristics, as, for instance, the growth of a plant that is shielded by a building wall, and the like.

Chapter 26

PRINCIPLES OF CLIMATOLOGICAL PROCESSING OF
MICROCLIMATE AND LOCAL CLIMATE OBSERVA-
TIONS -- THE UTILIZATION OF ULTRA-
SHORT SERIES OF OBSERVATIONS
AND THE CONSTRUCTION OF
LARGE SCALE MAPS

The specific characteristics of microclimate and local climate phenomena, distinguished by great variability in space, and also the short duration of observations of same, in the presence of general climatic data, determine the particular method of processing these observations.

In addition to the so-called primary processing, consisting of the interpolation of corrections, averaging of recurrences, basic tabulations and a critical check of same, the data collected on microclimate and local climate are to undergo special processing, which will provide for the maximum practical utilization of the results of observations.

No definite rules can be laid down for the processing procedure. Below, however, will be related the principles that are to be used as a basis, and examples, illustrating the application of these principles, will be cited.

The first phase in the processing procedure is the transition from the data of short-period observations to the so-called "normal" characteristics. Having obtained the "normal" character-

istic of an observation point, it becomes necessary to proceed to the evaluation of territories, which is the second phase of processing. The elementary method of linear interpolation is not acceptable in this case, and, for the characterization of territories with relation to local climate and microclimate, it is necessary to utilize the regularities, which tie the latter in with the basic features of a locality (terrestrial relief, vegetation, soil, structural development, etc.)

The methods of reduction to long series, used in general climatology, are adapted to observations, the periods for which are measured in years, while here we are dealing with ultra-short series of observations continuing only for several months, and, at times, for several days only. At the basis of the reduction of the ultrashort series of observations to the "norm" is the same method of differences or ratios. In the case of general climatology, the method is limited to deriving the mean difference (or ratio) for the period of one-time observations at the basic and the cited stations, proceeding from the tenet that the greater the number of cases, the greater the precision and the representativeness of the mean value derived.

To establish the characteristic difference, when only a small number of observations have been made, we cannot base ourselves on the law of large numbers, and must seek another method. It is here that the available data on the typical and characteristic general climatic conditions come to our aid, since the variation of the above-mentioned differences in time depends specifically on the changes in weather conditions (see Chapters 12 and 14 for the effect of sky cover and wind upon the features of microclimate and local climate).

The reduction to "norm" of ultrashort series of observations is done basically by calculating the regularities of the dependence of the contemplated differences upon the weather conditions, and by the probability computation of the above weather conditions, which are established over a long series of observations.

Such a reduction to "norm" was first utilized on a large scale in the processing of minimum temperature observations in the subtropical belt of the USSR. The best frost-hazard index for the territory, pertaining to the subtropical, all-year-round vegetating plants, is, according to G. T. Selyaninov, the mean value of the annual absolute minimum air temperatures. The normal index is computed by averaging the absolute annual minimums over a period of several decades, or by way of reduction over a series of years. As a result of specially organized minimum temperature observations for a series of points in Gruzija, diurnal temperature minimums were obtained for only two winter months of 1934. It was required, with the aid of these results, to derive an evaluation of the frost hazard of these locations, i. e., to derive the mean value of the absolute annual minimums.

In order to solve this problem, the regularities established by the analysis of the short-period observation data were utilized as follows. First of all, the quantitative ratio between the diurnal minimum temperature difference and cloudiness was established. Table 121 cites as an example the difference between the minimum temperatures of the upper and lower stations in Anaseul'. They are given in their mean values for all days (mean), for

non-overcast days (n.-o.), i. e., when the total value of cloudiness for the preceding evening and following morning measured 18 balls, and for overcast days (o.), i. e., when the above total value of cloudiness measured more than 18 balls. The Table shows that the minimum temperature differences depend not only on cloudiness, but that the fluctuations from year to year of the mean differences for the non-overcast days are smaller than the mean value for all days. Therefore, the accuracy of the first, regardless of the smaller number of cases than the mean for all days, is greater than the accuracy of the last.

Table 121

Mean minimum temperature differences

Years	Upper station - January			Lower station - February		
	Mean	n.-o.	o.	Mean	n. o.	o.
1930 - 1931	4.3	5.9	2.3	3.8	5.0	1.9
1931 - 1932	2.4	5.3	0.9	1.7	5.0	1.0
1932 - 1933	3.3	5.9	1.5	3.3	5.0	1.3
1933 - 1934	4.0	5.8	1.3	2.8	5.0	0.8
1934 - 1935	3.5	6.5	0.5	3.6	5.2	0.6

By the data of long-series observation points, the recurrence of cloudiness during absolute annual minimum temperatures was established as follows (Table 122).

Table 122

Recurrence of non-overcast and overcast weather
during absolute annual minimum temperatures
(in %)

Name of station	Number of years	Non- overcast	Overcast
Batumi	31	61	39
Poti	26	77	23
Samtredi	27	78	22
Makharadze	29	86	12
Sakari	41	78	22
Sukhumi	30	70	30
Sochi	27	81	19
Mean	--	76	24

From this, the deduction was drawn that the mean difference Δt for the mean values of the absolute annual minimums may be computed as a mean weighted value:

$$\Delta t = \frac{n \Delta t_{n-o.} + m \Delta t_{o.}}{n + m} \quad (36)$$

Where n is the percent of years with absolute minimum, which was accompanied by non-overcast in the given area; m is the same, but with overcast; $\Delta t_{n-o.}$ is the difference in the minimum temperatures under non-overcast skies; $\Delta t_{o.}$ is the same as above, but under overcast skies.

In order to compute the mean value of the absolute annual minimums for the contemplated observation points, the above-obtained difference was added to the mean value of the absolute annual minimums for the base station.

In the above-indicated example it was assumed that Δt varies only in relation to cloudiness. It is necessary to ascertain the magnitude of the error resulting from this assumption. The checking may be accomplished empirically, utilizing pairs of the long series stations and comparing the differences of the mean absolute minimums, obtained by direct juxtaposition of the absolute minimums with the difference obtained by formula (36).

In this case the check confirmed the correctness of the assumption. Otherwise, it would have been necessary to make a supplementary analysis and introduce corresponding corrections.

Such a method of reduction may be utilized for any element of microclimatic and local climate.

In a number of cases, it is not only the final total -- the mean weighted difference, -- that is of practical importance, but also its component parts, for which purpose it is necessary to know the probabilities of various differences.

If it is not only the weather conditions that vary in time, but also the elements of the landscape tied in with the weather, such as the height and the denseness of the grass cover, the latter are also to be taken into account in the reduction to the "norm".

In an emergency assignment, when it is necessary to complete the work in one season, this circumstance is to be taken into account in the working out of the schedule, providing for, regardless of weather conditions, the various character of the grass cover, utilizing sprinkling and other methods of agricultural technique, and passing thereby from simple observations to the arrangement of field tests.

Reduction to "norm" is also possible in the case of field observations along predetermined routes-profiles in those cases when we are dealing with the quantitative variation of the same phenomenon (for instance, the temperature distribution along a valley slope during the period of the temperature inversion), provided that, at the points of extreme values of the characterized element, systematic observations are made. The latter may be substituted by points with systematic observations under similarly extreme conditions. In both cases, the "normal" differences between the extreme points are taken and compared with the differences obtained for the same points by field observations. If their ratio is equal to unity, i. e., if the differences are equal, it is assumed that all the differences between the intermediate points of the course and one of the base stations also correspond to the "normal". If, however, the indicated ratio does not equal unity, then, in order to reduce to "norm", all other differences are multiplied by a factor, which is equal to the indicated ratio.

When passing from the data pertaining to an observation point to the characterization of a territory, regardless of

whether it is given in the form of maps or Tables, it is necessary to establish the regularities, linking the characteristics of microclimate or local climate with the physico^vgeographical conditions, by which they are stipulated. In connection with this, it becomes necessary to classify these conditions by their effect upon microclimate and local climate.

It is particularly difficult to lay out large-scale maps, in which an unbroken characterization of a territory must be presented, regardless of the complexity of physico-geographical conditions of its individual parts.

As an example, we will cite the construction of a map of the mean absolute annual minimum air temperatures and a soil-temperature map.

The first large-scale map of the mean absolute annual minimum temperatures, constructed on the above-indicated principles, was accomplished in 1933. It embraced the subtropical belt of Western Gruzija. A part of this map was presented in Figure 59 of the preceding text.

In the analysis of the regularities of the distribution of the contemplated element, the effect of the latitude and the altitude of the locality, the forms of terrestrial relief, and the proximity of the sea, were all taken into account, with the full utilization of not only the direct data available at meteorological stations, but also the data contained in the existing synoptic maps of days with severe frosts. It was possible to disregard the effect of latitude, with the exception of the sec-

tion from Adler to Sochi, since the latitude of the contemplated territory changes within the range of two degrees only. In addition, the synoptic analysis ascertained that the basic cold penetrations into Western Transcaucasia come not from the north, which is protected by the main mountain ridge, but from the west.

Thus, the analysis of the regularities over most of the territory could be reduced to the effect of three factors: proximity to the sea, terrestrial configuration, and altitude above sea level. For the altitude, an adiabatic gradient of saturated air in the value of 0.5 degrees per 100 meters, was assumed. Since the cartographing was done to a height of 500 meters only, an error in the gradient to the extent of 20 percent, i. e., to the extent of 0.1 degrees, would in the worst case, result in a cumulative error of 0.5 degrees, which is of no essential significance when the isotherms are plotted at 2-degree intervals.

To determine the effect of terrestrial configuration, three types of relief were segregated: (1) the upper parts of the slopes and the convex summits, (2) the plains, flat summits and well-ventilated valleys, (3) valleys and synclines.

All the available stations were classified by these three types of relief, the data for which were reduced to sea level, and for each type of relief a curve, characterizing the variation of the mean value of the absolute annual minimum temperatures in relation to the distance from the sea, was plotted (see Figure 84).

By the data of these graphs and by the altitude gradients, auxiliary tabulations of the mean values of the absolute annual minimums, as functions of three variables, were made, the three variables being: terrestrial relief, distance from the sea, and altitude above sea level. With the aid of these auxiliary tabulations, the map was constructed to a scale of 1:210,000 (5-"verst" scale), on the hypsometric principle, with horizontals at intervals of 25 and 50 meters. For the characterization of terrestrial relief, one-"verst" maps and direct knowledge of the territory were additionally utilized.

In a similar fashion, were constructed maps of the mean values of the absolute annual minimum temperatures for individual economies in the subtropical belt, to a scale of 1:10,000 (see Figure 85).

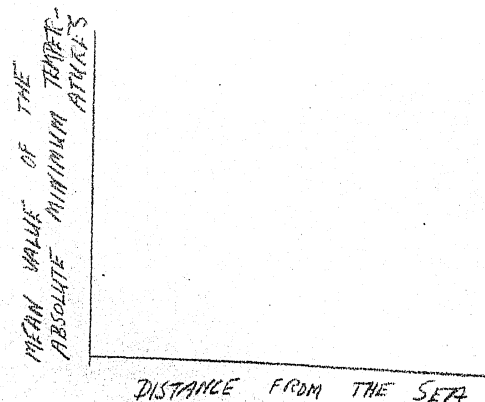


Figure 84. Variation of the mean values of the absolute annual minimum temperatures with distance from the sea -- for three types of terrestrial relief. (1) upper parts of slopes and convex summits; (2) plains, flat summits, and ventilated valleys; (3) valleys and synclines.

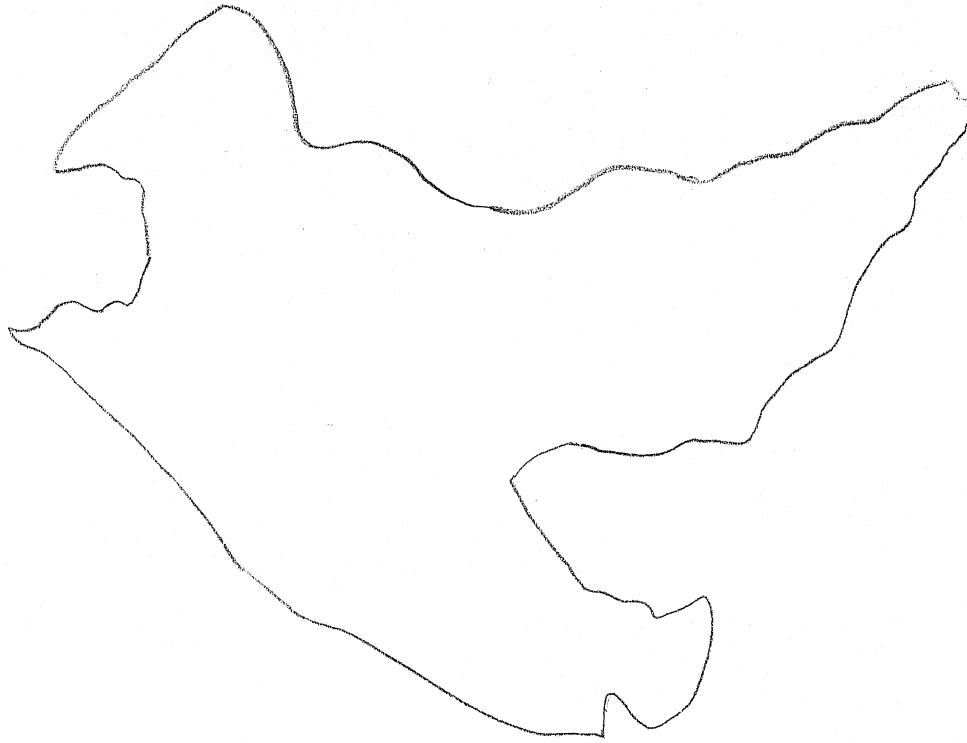


Figure 85. Detail zoning of Sovkhoz "Il'yich" by the degree of frost hazard.

An experimental attempt to construct a large-scale map of the soil temperature at the depth of 20 centimeters was made at the Agro-Hydro-Meteorological Institute by S. A. Mikhaylovskaya for the territory of a tea economy [plantation] in Western Gruzziya. The entire area was a hill-interlaced terrain with, approximately, a 70-meter altitude differential and a predominant slope steepness of 10-20 degrees. The assignment was somewhat facilitated by the fact that the entire territory was occupied by a single crop -- the tea plant, with a uniform agrotechnique being used. The soil-temperature observations were conducted in September-October 1936 at 15 observation points, located on slopes

of various exposure and of various altitudes from the foot of the slope. The assignment was to construct a temperature-distribution map for the period of observations.

Since the differences in the soil temperature of different slopes are determined by the various inflow of direct solar radiation, the ratio of which to the slope exposure is known, it is natural to, first of all, juxtapose the temperature of the soil with direct solar radiation. Figure 86 below represents a correlation graph of the mean diurnal soil temperatures by the diurnal totals of direct solar radiation, computed by A. N. Gordov for latitude 42 degrees north.

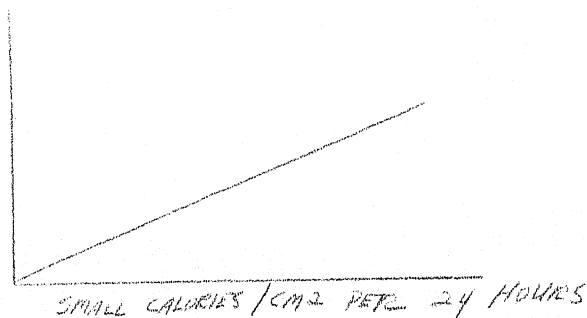


Figure 86. Mean diurnal temperature of soil at the depth of 20 centimeters in relation to diurnal totals of direct solar radiation.

The correlation factor for the soil temperature and direct solar radiation is 0.96. The two points, encircled by a dotted line in Figure 86, were not utilized in the computations, since the temperature of the soil under these points, was reduced at the expense of the greater denseness of the tea plants, which was

not representative for the greater part of the territory. The corresponding regression equation is as follows: $\Delta t = 0.014 \Delta Q$, where Δt is the temperature variation corresponding to the variation in the direct solar radiation ΔQ .

Taking into account the fact that the inflow of direct solar radiation is in direct ratio to the exposure and steepness of the slopes, it becomes possible, by the use of the regression equation, to tie in the temperature of the soil with the exposition of the slopes. This tie-in is represented in the form of isopleths in Figure 87.

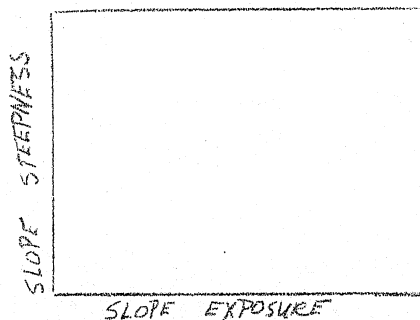


Figure 87. Mean diurnal temperature of the soil at the depth of 20 centimeters in relation to the exposure and steepness of the slopes (for the period September-October 1936).

The obtaining of the above ratio permits the construction of soil-temperature maps of a contemplated territory, by basing oneself on the hypsometric map. To accomplish this, it is necessary:

(1) To determine, for a series of points, chosen on the map, the exposure and steepness of slope.

(2) With the aid of the isopleths, to determine the temperature for the given exposure and steepness.

(3) To draw on the map a system of isolines.

The construction of a soil-temperature map is of particular interest, since it is not only the results of direct observations, but also theoretical computations that are being used in the process.

In processing observation data, pertaining to microclimate and local climate, a particular part is played by the analysis of the purely physical essence of phenomena, i. e., their genesis. As seen from the preceding argumentations, the outlining of the genesis of phenomena is absolutely necessary in the process of reduction to the "norm" and in the process of cartographing the data obtained. It is only by understanding the physical essence of the phenomena involved that the elements of weather and landscape on which the latter are predicated can best be segregated and underscored.

Of still greater importance is the thorough understanding of the genesis of the phenomena, when the assigned purpose of the investigations is the recommendation of measures to be undertaken, with a view to the changing of certain features of the microclimate and the local climate.

In such cases, the proper evaluation of the part played by individual components of the thermal balance, and also turbulence, as one of its factors, easily amenable to change, assumes a predominant importance.

APPENDIX I

METHODICAL INSTRUCTIONS FOR
ANEMOMETRIC SURVEYS

When organizing an anemometric survey, it is necessary to take into account the already existing data on the general regularities in the deformation of the air current by terrestrial relief and by the shielding afforded a locality by vegetation and buildings, as already analyzed in Chapters 6 and 12.

It is necessary to have a plan of the terrain involved to a scale 1:2,000 or 1: 10,000, depending on the degree of interlacing, with horizontal sections at intervals of 2-5 meters. Vegetation and structures are to be denoted by conventional markings. In order to construct a plan to such a large scale, a topographic map, 1:25,000, may be utilized, by correspondingly enlarging and precisioning it. To introduce more precision into the topographic map, particularly when dealing with interlaced relief, leveling is recommended for the profiles intended for the anemometric survey.

All points and routes of the anemometric survey are to be marked in the plan.

One of the important things in organizing a survey is the selection of a check point at an exposed level place. The importance of the check point is particularly great in those cases, when the entire area under survey cannot be embraced by simultaneous observations, and the survey is conducted in parts, i. e., certain sections are surveyed on certain days, while other sections are done on different days.

If the territory being surveyed embraces an area of several tens of square kilometers, it is useful to establish supplementary check points in direct proximity to the routes of the survey.

In the selection of a check point, it is necessary to be guided by the considerations below.

A check point is to occupy a spot of intermediate altitude, i. e., it must not be located at the summit of an elevation or at the bottom of a valley. The slope at the check parcel is to be less than 2 degrees for an area with a radius of 100 meters, it being the case that the distance to individual elevations, to a precipice, to vegetation, to structures is to be no smaller than 20 times their height, respectively. Individual structures and trees, which shut out the horizon to the extent of one compass point only, may be overlooked.

All other conditions being equal, more elevated locations are preferable. Wherever possible, the check point is to be adapted to an existing meteorological station.

As a rule, parcels with a diurnal shift in the direction of the wind are not suitable for check points. An exception may be made only in the case of territories, which are completely embraced by such shifts in the winds.

Individual points marked out for the survey are aligned by definite routes in a manner that makes it possible to trace the configuration of the air current deformation, with one ob-

server in charge of two points, when the distance between them is not in excess of 100-150 meters. When 5 observers are available, one check point and 8 route points can be serviced. The points of the route are preselected, in keeping with the assignment, confirmed after a test survey, fixed with precision by a stake driven into the ground or a stone, and numbered in the plan. Every observation point is assigned a double number: the number of the route (a Roman numeral) and the sequence number of the point within the route (an Arabic numeral). The location around each observation point is described.

At the check point, the following is done:

- (1) systematic observations of the wind and the vertical temperature gradient of the surface air layer (for the analysis of atmospheric stability), the observations being timed to the conventional four hours (0100, 0700, 1300, and 1900 hours);

- (2) continuous episodical observations of the wind and the vertical temperature gradient directly at the time of the anemometric survey.

For systematic observations, the Wilde weather vane is used, whenever possible, on a pole not lower than 6 meters, for purposes of comparison with general meteorological observations, and anemometers mounted at the same height as in the survey.

Determination of the vertical temperature gradient is done with the aid of ventilation psychrometers at the heights of 20 and 150 centimeters.

The equipment of the anemometric survey observation points consists of a number of anemometers corresponding to the number of heights, at which readings are to be taken. The direction of the wind is determined by the anemometer pennants. For orientation by the points of the horizon, compasses are used, but it is desirable to have at each observation point local point-of-horizon references (for instance, due south -- the chimney of a plant; due southeast -- the angle of a fence; and the like), or reference peg marks.

Timepieces are to be available for the fixation of the starting and stopping moments for the anemometers. When distances are small, time may be signaled from one point to the next.

Anemometers are mounted on poles. Poles above two meters in height are braced on three guys, with the bottom end of the pole just resting on the ground. To mount the anemometer, the pole, without the guys' being loosened, is lifted slightly, and its bottom end moved to the side. Over the top end of the pole there is a braced capping, into which is fitted a plug with the anemometer screwed on over it. The starting and the stopping of the anemometers at heights of over 2 meters is effected by a starting cord fastened to the corresponding lever of the device.

The timing of the anemometric survey is established in relation to the assigned task. The duration of each series of observations is about 1.5 hours, for which time interval it is possible to accomplish several observations under uniform weather conditions. In selecting the time for observations, transition

hours, as indicated in Table 8 of the preceding text, are to be avoided, unless they are of self-contained interest (for instance, the determination of the time of the discontinuation of the nocturnal breeze).

All observers converge before observation time to the check point, where they check their timepieces and anemometers (by running them for 10 minutes), then go back to their individual points timing themselves so that at a predetermined moment they will all be in their places ready to proceed with their observations. Observations are usually conducted by a repeat start of the anemometers, each time for a period of 10 minutes.

The starting and the stopping of the instruments are to be synchronized for all points of observation. In the case when one observer services two points, the starting and the stopping of the instruments at the second point may follow the same at the first point by 1-2 minutes, and it must be so noted in the observation log book. The direction of the wind is determined while the meters are at work. The anemometer recordings are taken during the 5-minute interval between the stopping and the next starting of the meters.

At the same time, observations are proceeding at the check point with supplementary psychrometer recordings (may be done without sprinkling the thermometer) at the heights of 20, 150, and, again, 20 centimeters. In addition, the amount of cloudiness and of hydrometeors is recorded, and the Wilde weather vane, when present, is checked off.

In addition to the parallel control checks upon the anemometers before the survey is started, they are also subject to special checks at least every ten days. For this purpose, the anemometers to be checked are set up on a horizontally braced rod, at a height of 1.5-2.0 meters. The rod is turned each time normal to the direction of the wind. A specially selected hand-operated anemometer is used as a check in this particular procedure. Six-eight anemometers being checked are started simultaneously, together with the control anemometer, 3 times for a 10-minute period each time. The next check is to be timed to other wind velocities. The results of the check are carried on to the calibration graph, which is plotted by the certificate of the instrument being checked (Figure 88), with the readings of the check anemometer laid off along the Y-axis, and the readings of the anemometer being checked -- along the X-axis.

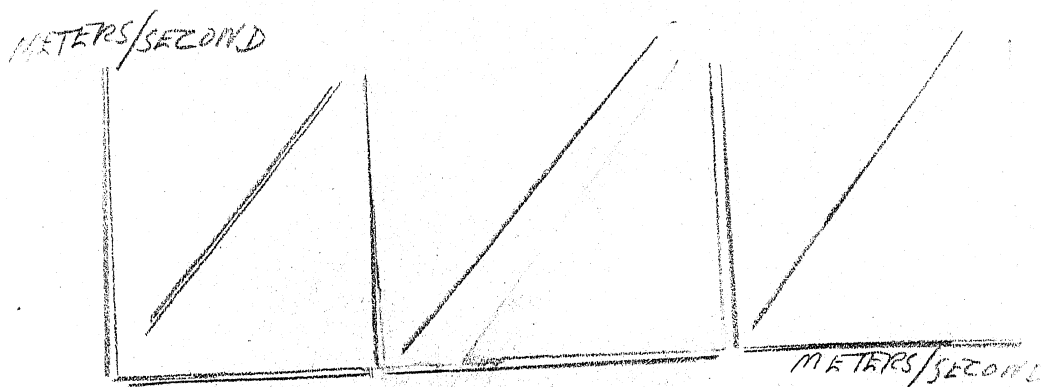


Figure 88. Results of checking anemometers under field conditions. Various markings designate results of checks on different days.

Three examples of such graphs are shown in Figure 88. The first example (A) shows the results of checking an anemometer, which is in good condition. The maximum departures of individual observations from the calibration curve are not in excess of 0.3 meters/second. In the second example (B), greater departures can be seen, exceeding in some cases 0.5 meters/second. Such anemometers are to be rejected. Finally, in the third example (C), are shown check results that permit the plotting of a new calibration curve (see dotted line).

The departure of individual check points from the calibration curve may be considered as the index of the accidental error value. Under the accepted rejection principle, accidental errors of anemometers amount to approximately ± 0.1 meters/second in their mean value. The variations in the calibration curve, which were sometimes gradual and sometimes abrupt, and which could not always be taken into account in good time, are to be considered as the characterization of the possible systematic error of the individual device.

Experience shows that, with the special selectivity of instruments and their continuous checking, a systematic error of 0.1 meters/second can be attained, whence a total error of ± 0.2 meters/second is derived.

The results of the survey (wind velocity, converted into meters/second by the corresponding certificates, and wind direction) are carried over into a Table, into which are also carried the ratios of the wind velocities of each observation point to the synchronous wind velocity at the check point, and also the

respective differences in the direction of the wind.

Next, the data are grouped by wind velocities and directions, also by atmospheric stability, the corresponding mean values are computed, and, finally, compound Tables of the type shown in Table 123, are constructed.

Table 123

Variation in the nocturnal wind velocity at
the height of 1 meter, in relation to
the form of terrestrial relief,
expressed in ratios to the
wind velocity over an
exposed level spot

Form of terrestrial relief	At a wind velocity of	
	2 meters/sec	2-4 meters/sec
Summit of steep elevation	1.2 (3)	1.6 (3)
Summit of shallow elevation	1.9 (7)	1.2 (6)
Hillock	1.4 (4)	1.4 (5)
Windward and top	1.8 (2)	1.6 (4)
parallel to the center	1.6 (2)	1.2 (4)
wind slope bottom	1.4 (2)	1.0 (4)
Leeward slope top	1.6 (4)	1.1 (3)
center	1.1 (4)	0.7 (2)
bottom	0.9 (4)	0.5 (2)

Note: The number of observations is shown in parentheses.

In the case of wind velocities of less than 1.0 meters/second, a hand-operated anemometer is not suitable for observations. During the daylight hours, such winds are usually of directional instability and are of no particular importance. At night, however, local winds of this specific velocity are of great directional stability (downslope, along the valley). The study of the velocity and direction of these winds is of great practical importance.

The most widely used method for the fixation of such winds, predominantly their directions, is the method of smoke screens. To produce the smoke, bonfires or special smoke-producing logs are used. The direction of drift of the smoke is ascertained with the aid of a compass. At spots, where a possible change in the drift direction of the smoke may occur, additional observers are stationed for the purpose of fixating the changed directions. To fixate the point of turning of the smoke screen and carry it over to the map, it is necessary to measure not only the directions, but also the distances between the turning points, which is accomplished very simply by the use of the two-meter rod ("sazhenka").

To determine the direction and velocity of a wind blowing at less than 2 meters/second, and also to determine the air stagnation zones under conditions of a temperature inversion and a state of atmospheric equilibrium, a moored balanced pilot balloon can be used. Since this method was not published before, we will analyze it in greater detail. It is necessary to

have the following equipment: (1) a No. 5-15 balloon envelope, (2) hydrogen, (3) a small cloth bag having a volume of 200 cubic centimeters, (4) tough cordage, (5) a plane table with compass, (6) a compass, (7) a sighting rule, (8) a chronometer, (9) a two-meter rod ("sazhenka").

It takes 3 observers to conduct these observations. Two of them manage the pilot balloon along established routes, the third one conducts parallel observations of wind velocity and direction at the check point. The observations consist in fixing on the plane table and recording in the observer's notebook of the direction and velocity of the pilot balloon, which moves freely together with the air current, at a height of 1.5-2 meters, at 1-minute intervals.

Parallel observations with a pilot balloon (No. 15 envelope) and an anemometer resulted in the balloon indicating velocities by 0.2-0.3 meters/second lower than the anemometer. This difference can be explained by the following: the balloon, the diameter of which ≈ 0.7 meters, sustains the effect of the velocities of vortices and currents of a larger scale, while the anemometer reacts to velocities of smaller vortices, regardless of their directions. This feature of the pilot balloons is even of some advantage, particularly in determining stagnation zones, the typical characteristic of which is the absence of a directed current.

For purposes of observation, the envelope of the pilot balloon is filled with hydrogen up to a lifting force of 70-100 grams, after which a tough cord 4-5 meters long is fastened to it. For purposes of balancing, a bag filled with sand is tied by a loop to the cord at a distance of 0.5 meters from the balloon (experience showed that one balloon can operate for several hours).

The 2-meter rod ("sazhenka") is knocked together from 2 sticks braced with a cross bar. The distance between the free ends of the sticks should be 2 meters.

Prior to observations, initial points are selected for the routes. It is rational to locate these points on the windward boundary of the contemplated parcel, along a line normal to the direction of the wind. The distance between the points is determined by the degree of non-uniformity of the parcel, and may fluctuate from 50 (for an interlaced locality) to 200 meters (for a level locality). In the selection of these points in an interlaced locality, it is necessary to be guided by the forms of relief, and, in the presence of vegetation, it is also necessary to take into account its characteristics. All the initial points are to be plotted on the plane table. A plan of the parcel, as surveyed roughly by the eye, is also to be marked out on the plane table.

The work is performed by two observers. The first one manages the pilot balloon, the compass, and records the observa-

tions in the field notebook. The second one operates the plane table with the compass, the sighting rule, the chronometer and the two-meter frame. The order of observations is as follows. Both observers stand themselves at the first initial point. The first observer marks down in the field book the date, the name of the locality, the numbers of the observation points, between which the measuring will be done, and the time (hours and minutes) of the beginning of observations. Ten seconds before the predetermined full minute, the second observer commands to "make ready", and, at the moment of the full minute, he shouts the command for the start. At this command, the first observer loosens the balloon cord to a length of 2-3 meters and follows the motion of the pilot balloon. The absence of tautness in the cord is the index of the free motion of the balloon. Upon the expiration of one minute, the second observer shouts "stop", and the first observer stops in his tracks. Having stopped, the first observer fixates the general direction in which he was moving, sighting by the compass the spot, where the second observer stands. The latter, in turn, sights and draws in on the plane table the direction on the first observer. After the direction has been fixated, the second observer measures with the aid of the 2-meter rod the distance to the first observer, and lays it off on the plane table, thereby having marked out the next observation point of the route. The distances are laid off to a scale of 1:2,000 or 1:5,000. The first observer records in his field book the directions of the balloon motion, the number of 2-meter rods it took to cover the distance, and their transla-

tion in terms of meters/second (by dividing by 60). By foot-note, he records the nature of the balloon's motion: departures from direction, non-uniformity of motion, also the characteristics of the locality. A change in the direction of the wind and the deflection from a straight line in the path of the balloon as a result of it, are marked on the plane table by a dotted line, and the features of the locality are marked in by conventional topographic designations. The determination of the direction in which the balloon moves, by both observers, serves as a mutual check.

After the recording of the first observations in the field book and the fixation of their results upon the plane table, observations are continued in the same order all the way to the boundary of the contemplated area, or to a parcel with a wind velocity of over 2 meters/second, or to the boundary of the air stagnation zone. The latter is determined by the absence of a directional motion (with a velocity of more than 0.1 meters/second) over a period of several minutes. Having completed the survey of the first route, the observers pass to the next initial point and begin the survey of the second route in the same order of sequence. Figure 89 is an example of the way results of observations are carried onto the plane table. In the case when it became necessary to halt the progress of the pilot balloon, or the balloon became entangled by the cord in the undergrowth, the first observer is to stop, and, upon returning to the preceding point, repeat the observation from the very beginning.

If the pilot balloon for some reason or other drifts higher or below the 1.5-2.0 meter altitude, it is necessary to either jerk it down or lift it by the thread to its proper level. The persistence of the balloon in rising above or descending below the predetermined level points to the necessity of re-balancing it. If along the path of the balloon there is an obstacle (a ditch or fencing), which will detain the observer, the minute observation interval is to be reduced. In such a case, the first observer, having reached the obstacle, which, while not hindering the movement of the balloon, will detain his own progress, signals a "stop", and the second observer records, in seconds, the time of the stop. The direction of the motion is determined in the usual manner, while its velocity is computed by dividing the distance already covered by the actual number of seconds the balloon was in motion until the stop. Upon the crossing of the obstacle by the observer, the observations are resumed.

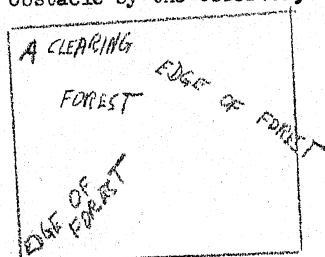


Figure 89. Example of a survey of the direction and velocity of air currents with the aid of a moored pilot balloon -- with the wind direction over the exposed location WSW-SSE, and its velocity 2.2 meters/second.

For the appropriate evaluation of the results of observations and an accounting of the wind variations in time during the period of the balloon observations, continuous observations of the velocity and direction of the wind are conducted at the base point for the 2-meter height. An anemometer is run during the entire period of observations, its recordings taken at 5-minute intervals, without stopping it. The direction of the wind is determined by the anemometer pennant and by the compass. The anemometer may be substituted by a Tret'yakov wind gauge, the recordings of which are also counted every 5 minutes. If a nearby meteorological station is utilized as the base point of the survey, the direction and the velocity of the wind are determined additionally by the weather vane (every 5 minutes).

The results of the pilot balloon observations are juxtaposed with the data of the base station.

APPENDIX 2

DESCRIPTION DIAGRAM OF THE
OBSERVATION POINT LOCATION

NAME OF POINT, COORDINATES,
ALTITUDE ABOVE SEA LEVEL

I. TERRESTRIAL RELIEF
AND ADJACENT BODIES OF WATER

1 - General Analysis

Level, hilly, or mountainous terrain. General direction and characteristics of surrounding elevations, the order of the altitudes. Direction and characteristics of the valley. Transverse profile of valley. Width of valley. Longitudinal profile of valley. Presence of sharp turns or constriction of valley (above or below the meteorological station).

With the proximity of a sea, a lake, a river, or any other body of water, a determination is to be made of the distance to it, the size of it, the direction of its location with reference to the meteorological station (by points of compass), the general direction of the shoreline (existing large-scale maps may be utilized), and the accessibility to it.

2 - The Analysis of the Observation Point

(a) If located on a slope:--Slope into valley, exposed slope toward the sea or toward plain; exposure, steepness of

slope in degrees at the instrument mountings, as well as below and above these; if located on a slope terrace, the terrace width is to be indicated. Altitude above the bottom of the valley or plain. General extent of slope in a horizontal direction. The part of the slope (upper, middle, lower), where observation point is located. Observation point at convex or at concave part of slope. When observation point is on slope, facing valley, the latter to be described in detail as per diagram I, 2(b) directly below.

(b) When observation point is at the bottom of a valley, the description of the valley as applicable to the observation spot itself (see I, 1) is made more precise. Exits of secondary valleys and ravines above and below the observation point are noted. The characteristics of the valley bottom, particularly the slopes, at which the observation ^{point} lies (see I, 2a).

The distance to the waterline at the river bank, the width of the river at flood stage and at mean water level .

(c) If located on a summit, - the nature of elevation is to be described, whether hill, individual mountain, ridge. The altitude above the bottom of valley or plain. The nature of the slopes (see I, 2a). Description of the summit itself if flat, the dimensions of the flat part, its shape (round, elongated, in which direction) to be indicated. The distance of the meteorological station from the edge of a horizontal platform, or from the edge of a precipice (in which direction).

II VEGETATION

1. General Description of Vegetation in the Observation Point Area.

Forest - deciduous, acerose; sparse, dense; low-height trees and undergrowth thickets; fields, orchards; inundation and dry meadowland; marshland, drained and undrained; wind-protection strips; and complete absence of vegetation (sands, denuded rocks, etc.). Location of individual vegetation masses with reference to observation point to be indicated. The presence of vegetation masses which may hinder the flowoff of air masses during nocturnal inversions.

2. Vegetation at Meteorological Platform and at Directly Adjacent Parcels.

Nature of the vegetation cover: artificial (black fallow, trampled-on), or naturally denuded soil; natural meadowland or steppe vegetation, cultivated crops. Height, denseness of grass cover, if possible, the composition of its species, the amount of vegetation mass per one square meter during various seasons.

The presence of undergrowth and trees, their species composition, height, degree of denseness, the number of tree trunks per 10 square meters, the presence of young forest growth, the distance, direction from individual trees. Wind-protection strips:- the direct proximity of distance from, structure of, height, species composition, ventilability of, and structural variation of its height.

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III - SOIL

Characteristics of the soil surface. Description of soil by genetic layers (the presence of sod). Physical properties of soil. Level of surface waters and its dynamics. In the presence of a snow cover, its depth, compactness, structure, and character of stratification to be described.

IV. SURROUNDING STRUCTURES

(1) Whether observation point located in city, or large village, at the outskirts (northern, eastern, and the like), in the center. Predominant character of structures (multiple-story, single-story, continuous building line, etc.).

(2) Height, dimensions, distances, and directions of all buildings surrounding the meteorological platform (to a distance 20 times the structural height).

(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)
VI	63	60	54	46	36	27	17	9	1	-
VII	63	60	53	45	36	26	16	8	0	-
VIII	53	52	46	39	30	20	11	2	-	-
IX	43	41	37	29	21	12	2	-	-	-
X	32	30	26	20	12	4	-	-	-	-
XI	22	20	17	12	5	-	-	-	-	-
XII	16	15	12	7	-	-	-	-	-	-

Latitude 55 degrees

I	14	13	10	5	-	-	-	-	-	-
II	22	21	18	13	6	-	-	-	-	-
III	33	31	28	22	15	7	-	-	-	-
IV	45	43	39	33	25	17	8	0	-	-
V	54	53	48	40	33	24	15	7	0	-
VI	58	56	51	44	36	27	19	11	4	-
VII	56	55	50	43	35	26	18	10	2	-
VIII	48	47	43	36	28	20	11	3	-	-
IX	40	38	34	26	19	11	3	-	-	-
X	27	26	22	17	10	2	-	-	-	-
XI	17	16	13	8	2	-	-	-	-	-
XII	11	10	7	3	-	-	-	-	-	-

Latitude 60 degrees

I	9	8	5	2	-	-	-	-	-	-
II	17	16	13	8	3	-	-	-	-	-
III	28	27	23	19	13	6	-	-	-	-

[continued on following page]

(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)
IV	40	38	35	30	23	16	8	2	-	-
V	49	47	44	38	31	23	16	9	3	-
VI	53	51	47	41	35	27	20	13	7	1
VII	52	50	46	41	33	26	19	12	5	0
VIII	44	42	39	33	26	20	12	5	-	-
IX	33	32	28	24	17	10	3	-	-	-
X	22	21	18	13	7	0	-	-	-	-
XI	12	11	8	4	-	-	-	-	-	-
XII	6	5	3	-	-	-	-	-	-	-

E N D

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